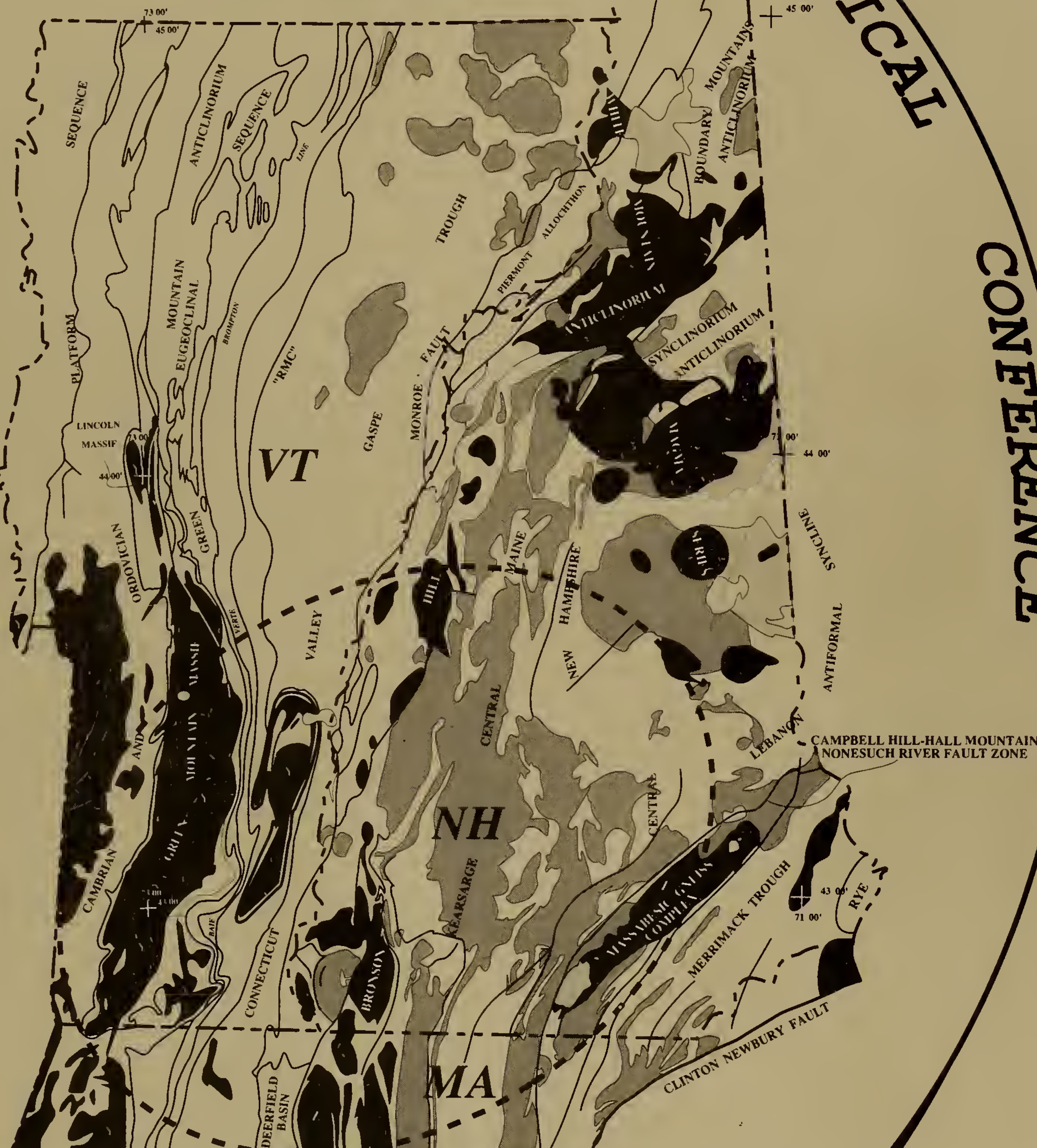
# NEW ENGLAND INTERCOLLEGIATE GEOLOGICAL CONFERENCE

## 1988

80th ANNUAL MEETING
OCTOBER 14 - 16, 1988
KEENE, NEW HAMPSHIRE

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NEW ENGLAND INTERCOLLEGIATE GEOLOGICAL CONFERENCE
*80th Annual Meeting*

# GUIDEBOOK FOR FIELD TRIPS IN SOUTHWESTERN NEW HAMPSHIRE, SOUTHEASTERN VERMONT, AND NORTH-CENTRAL MASSACHUSETTS

October 14, 15, and 16, 1988
Keene, New Hampshire

EDITOR

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Figure 1. Location map is an enlargement of the cover showing the areas covered by field trips. Dashed line boundaries - Friday trips; solid line boundaries - Saturday trips; dotted line boundaries - Sunday trips

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Cover: Geology generalized and digitized from Doll *et al.* (1961) and Stanley and Ratcliffe (1985) for Vermont; Lyons *et al.* (1986) for New Hampshire; and Zen (1983) for Massachusetts by J. Dykstra Eusden, Jr.

## FOREWORD: NEW ENGLAND -- 80 YEARS OF NEIGC

*"How shall we ever learn anything about orogeny, metamorphism, stratigraphy, paleoecology, or any other field of geology unless field investigations are encouraged?"* M. P. Billings (1950, GSA Bull. 61, p. 445).

After more than eight decades of informal field collegiality among hundreds....thousands....of university faculty, other professionals, and students from scores of departments in the greater New England area, it is appropriate to recognize the very significant advances that have been made in our geologic understanding of an admittedly small geographic area. The northern Appalachian Mountains are laden with exciting geologic problems, pure and applied, that 'run the gamut' in our increasingly diversified science. Those problems have generated hosts of solutions: some "unexpected", some welcomed, and some (many?) controversial. Each has taken us at least a small step forward, and provided a springboard for continued efforts. NEIGC has played a major role in promoting informal, constructive, even sometimes "fiery", discussions *on the outcrop* that in the final analysis has opened minds, honed observations, and refined ideas. We have learned from one another and almost always come away with a greater understanding of, and appreciation for, the problems at hand.

This year also celebrates the 20th anniversary of the publication of the Billings' Volume (Zen *et al.*, 1968), noted eloquently in this guide by Karabinos and Laird. That volume quickly became, and continues to be, a poignant starting point for many ongoing studies in New England. Work summarized there spurred continued field studies that have resulted in major compilations including four new state geologic maps (Massachusetts, Zen, 1983; Connecticut, Rodgers, 1985; Maine, Osberg *et al.* 1985; New Hampshire, Lyons *et al.*, 1986) with both Rhode Island and Vermont maps underway, regional tectonic syntheses (Williams, 1978; Stanley and Ratcliffe, 1985; Lyons *et al.*, 1982; Hall and Robinson, 1982), and the DNAG publications (eg., Roy, 1987; Thompson *et al.*, 1986) covering both the on- and offshore regions of the Northern Appalachians. As they should, each begins and ends *in the field.*

Field trips for this year concentrate on the area in southwestern and central New Hampshire, southeastern Vermont, and north-central Massachussets (figs. 1 and 2) which has not been visited by NEIGC for more than 20 years. The area records, in a highly compressed fashion, the effects of at least one billion years of earth history from Precambrian Y to the Quaternary. Ratcliffe *et al.* (Trips A-1 and B-1) and Karabinos and Laird (Trip C-3) emphasize the Grenville basement and its cover sequences, the effects of both the Taconic and Acadian orogenies, and the present best estimates for timing. Rosenfeld *et al.* (Trip B-6) provide new and exciting isotopic data from "snowball" garnets that indicate possible pre-Taconic growth as well as the rate of growth during Acadian polydeformation.

Stratigraphic, structural, metamorphic, and a bit of igneous geology are well represented in the Connecticut Valley trough (Hatch, 1987), the Bronson Hill anticlinorium, and the Central Maine Terrain (Zen, in press) and are attributed to Acadian activity. Hepburn (Trip B-8) updates work done in the Brattleboro quadrangle, Vermont, and discusses the important, and controversial, boundary between the "Vermont sequence" and the "New Hampshire sequence" while Schneiderman (Trip B-5) provides an opportunity to examine evidence preserved at Little Ascutney Mountain that the New Hampshire sequence overlay the Vermont sequence during Mesozoic White Mountain magmatic activity. The earlier and recently revived "probable late Ordovician age" for the Vermont sequence based on graptolites (Bothner and Finney, 1986), however, was discounted this summer when we re-examined the *outcrops*, and more lower Devonian plants were recovered from the Gile Mountain equivalent in Canada (Hueber *et al.*, in prep).

Acadian nappe formation, thrusting, and subsequent doming along the trace of the Bronson Hill anticlinorium is reviewed by those who originally identified them and, with their students, refined them. J. B. Thompson, Jr. (Trip C-3) reviews the Skitchewaug nappe in one of several areas where fossil control is "abundant" at high metamorphic grade. Chamberlain *et al.* (Trip A-2) carry the Skitchewaug nappe farther south and examine it and its nested neighbor, the Fall Mountain nappe. Robinson (Trip C-4) and his recent students, David Elbert (Trip A-5) and Peter Thompson (Trips B-2 and C-1) look at the evidence for early Acadian thrusting and the evolution of the Bernardston nappe from northern Massachusetts to the spectacular exposures on classic Mount Monadnock.

To the east Lyons (Trip A-4), Duke *et al.* (Trip C-5), and Eusden (Trip A-3) traverse the Central Maine Terrain in New Hampshire (Kearsarge-Central Maine synclinorium, Central New Hampshire anticlinorium, and Lebanon synformal anticline) and shed new light on the metamorphic and igneous geology of this important Acadian region. Recent recognition of the Rangeley (Maine) stratigraphy of Osberg *et al.* (1968) and Moench and Boudette

(1970) across central New England (Hatch *et al.*, 1984; Eusden *et al.*, 1986) has led to a clearer picture of the internal structure of the previously defined Merrimack synclinorium of Billings (1956). Detailed studies of Rumble and Chamberlain (Trip B-7) involving the role of fluids during metamorphism and magmatism revolve around historically important and no longer so enigmatic graphite deposits. The Acadian seems better understood as the Alleghenian emerges farther west (eg., Hepburn, and Eusden, this volume).

Well beyond academic interest, the increasing importance of Mesozoic and Cenozoic geology to "public needs" is reflected in the remaining five trips. A thorough analysis by Wise (Trip C-7) of brittle fracture phenomena in the Deerfield Basin as stress recorders has as important applications to present seismicity as it does to reconstruction of Mesozoic spreading. "Societal need to know" the distribution of surficial deposits, particularly the glacio-fluvial, has spurred renewed efforts to unravel the deglaciation history of New England. The Connecticut River valley (Larsen and Koteff, Trip A-6) and its tributary valleys (Washington and Larson, Trip B-3; Ridge, Trip B-4) preserve an unusually good record of deglaciation through the analysis of morphosequences and of the geomorphic development of this area (Caldwell, Trip C-6). Much of our future planning depends heavily on a continued commitment here and throughout the northeast.

It is clear that NEIGC serves as an important vehicle to respond to BIllings' 1950 question. This informal, loosely-knit organization also raises an annual challenge -- to editor and to author -- to provide a platform for healthy debate, for learning, and for comradery. As our "permanent secretary" D. W. Caldwell noted (1970, p. ii): "Throughout its history the sole purpose of the NEIGC has been to bring together in the field those geologists interested and active in New England Geology, to consider and discuss the results of new mapping and other geologic studies." The tradition continues!

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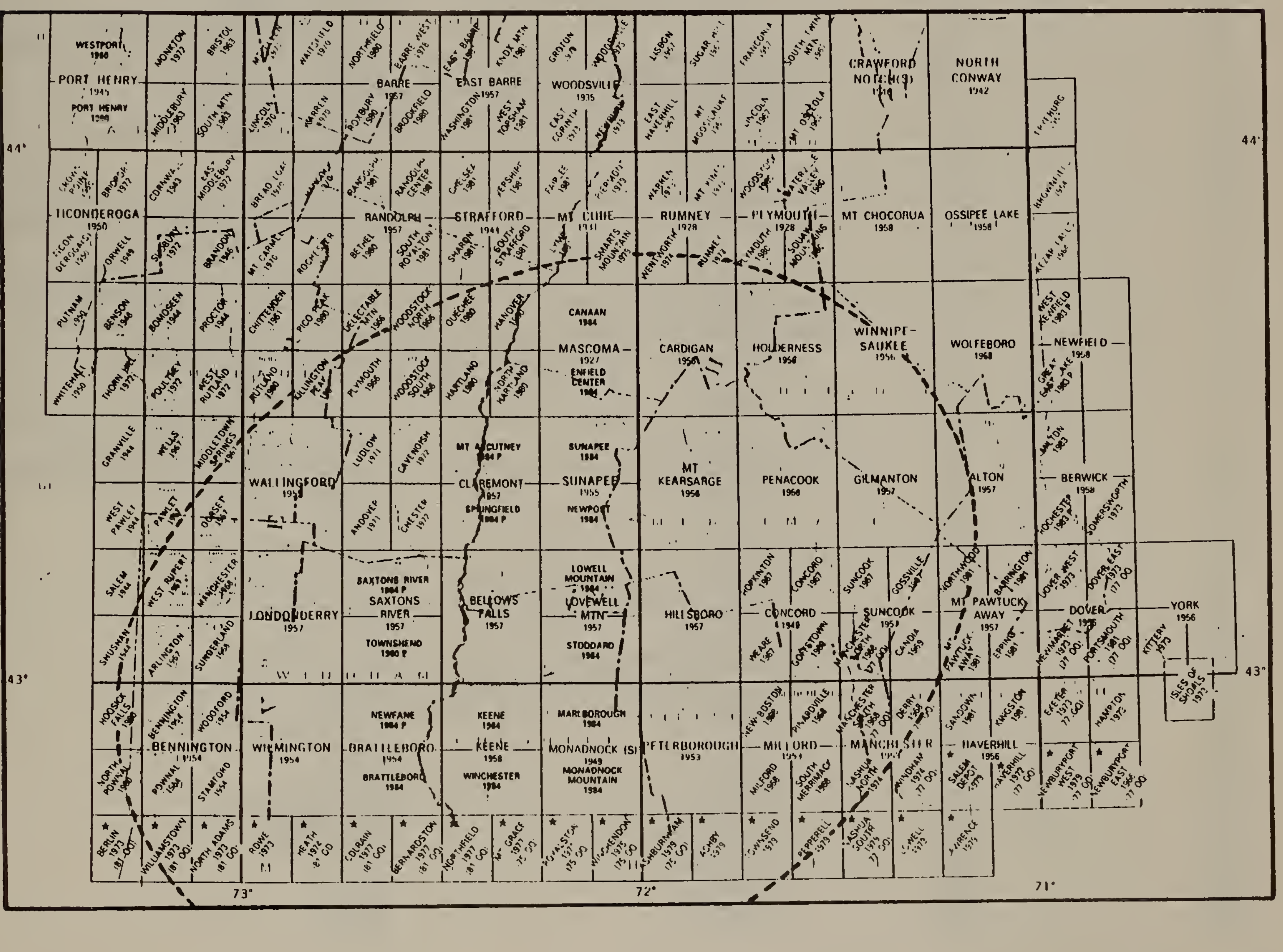
## ACKNOWLEDGEMENTS

I am very grateful for the willing and able assistance of Linda K. Hoadley and Natalie Simoneau in managing the "affairs of state" of NEIGC for 1988. Without that help, we would surely have been innundated, buried, and lithified.

Wallace A. Bothner, Editor
September, 1988

Figure 2. Topographic map index for southern New Hampshire, southern Vermont, and northern Massachusetts. After U.S. Geologic Survey Topographic Index (1986) for New Hampshire and Vermont.



CHRONOLOGICAL LISTING OF MEETINGS OF THE
NEW ENGLAND INTERCOLLEGIATE GEOLOGICAL CONFERENCE

| Meeting | Year | Location | Organizer |
|---|---|---|---|
| 1st | 1901 | Westfield River Terrace, MA | Davis |
| 2nd | 1902 | Mount Tom, MA | Emerson |
| 3rd | 1903 | West Peak, Meriden, CT | Rice |
| 4th | 1904 | Worcester, MA | Emerson |
| 5th | 1905 | Boston Harbor and Nantasket, MA | Johnson, Crosby |
| 6th | 1906 | Meriden to East Berlin, CT | Gregory |
| 7th | 1907 | Providence, RI | Brown |
| 8th | 1908 | Long Island, NY | Barrell |
| 9th | 1909 | North Berkshire Mountains, MA | Crosby, Warren |
| 10th | 1910 | Hanover, NH | Goldthwaite |
| 11th | 1911 | Nahant and Medford, MA | Lane, Johnson |
| 12th | 1912 | Higby-Lamentation Blocks | Rice |
| 13th | 1915 | Waterbury to Winsted, CT | Barrell |
| 14th | 1916 | Blue Hills, MA | Crosby, Warren |
| 15th | 1917 | Gay Head and Martha's Vineyard | Woodworth, Wigglesworth |
| 16th | 1920 | Lamentation and Hanging Hills | Rice, Foye |
| 17th | 1921 | Attleboro, MA | Woodworth |
| 18th | 1922 | Amherst, MA | Antevs |
| 19th | 1923 | Beverly, MA | Lane |
| 20th | 1924 | Providence, RI | Brown |
| 21st | 1925 | Waterville, ME | Perkins |
| 22nd | 1926 | New Haven, CT | Longwell |
| 23rd | 1927 | Worcester, MA | Perry, Little, Gordon |
| 24th | 1928 | Cambridge, MA | Billings, Bryan, Mather |
| 25th | 1929 | Littleton, NH | Crosby |
| 26th | 1930 | Amherst, MA | Loomis, Gordon |
| 27th | 1931 | Montreal, PQ | O'Neill, Graham, Glark, Gill, Osbome, McGerrigle |
| 28th | 1932 | Providence-Newport, RI | Brown |
| 29th | 1933 | Williamstown, MA | Cleland, Perry, Knopf |
| 30th | 1934 | Lewiston, ME | Fisher, Perkins |
| 31st | 1935 | Boston, MA | Morris, Pearsall, Whitehead |
| 32nd | 1936 | Littleton, NH | Billings, Hadley, Cleaves, Williams |
| 33rd | 1937 | New York City-Duchess Co., NY | O'Connell, Kay, Fluhr, Hubert, Balk |
| 34th | 1938 | Rutland, VT | Bain |
| 35th | 1939 | Harford, CT - Conn. Valley | Troxell, Flint, Longwell, Peoples, Wheeler |
| 36th | 1940 | Hanover, NH | Goldthwaite, Denny, Shaub, Hadley, Bannerman, Stoiber |
| 37th | 1941 | Northampton, MA | Balk, Jahns, Lochman, Shaub, Willard |
| 38th | 1946 | Mt. Washington, NH | Billings |
| 39th | 1947 | Providence, RI | Quinn |
| 40th | 1948 | Burlington, VT | Doll |
| 41st | 1949 | Boston, MA | Nichols, Billings, Shrock, Currier, Steams |
| 42nd | 1950 | Bangor, ME | Trefethen, Raisz |
| 43rd | 1951 | Worcester, MA | Lougee, Little |
| 44th | 1952 | Williamstown, MA | Perry, Foote, McFadyen, Ramsdell |
| 45th | 1953 | Hartford, CT | Flint, Gates, Peoples, Cushman, Aitken, Rodgers, Troxell |
| 46th | 1954 | Hanover, NH | Elston, Washburn, Lyons, McKinstry, Stoiber, McNair, Thompson |
| 47th | 1955 | Ticonderoga, NY | Rodgers, Walton, MacClintock, Bartolome |
| 48th | 1956 | Portsmouth, NH | Novotny, Billings, Chapman, Bradley, Freedman, Stewart |
| 49th | 1957 | Amherst, MA | Bain, Johannson, RIce, Stobbe, Woodland, Brophy, Kierstead, Webb, Shaub, Nelson |
| 50th | 1958 | Middletown, CT | Rosenfeld, Eaton, Sanders, Porter, |
| 51st | 1959 | Rutland, VT | Zen, Kay, Welby, Bain, Theokritoff, Osberg, Shumaker, Berry, Thompson |
| 52nd | 1960 | Rumford, ME | Griscom, Milton, Wolfe, Caldwell, Peacor |
| 53rd | 1961 | Montpelier, VT | Doll, Cady, White, Chidester, Matthews, Nichols, Baldwin, Stewart, Dennis |
| 54th | 1962 | Montreal, PQ | Gill, Clark, Kranck, Stevenson, Steam, Elson, Eakins, Gold |
| 55th | 1963 | Providence, RI | Quinn, Mutch, Shafer, Agron, Chapple, Feiniger, Hall |
| 56th | 1964 | Chestnut Hill, MA | Skehan |
| 57th | 1965 | Brunswick, ME | Hussey |
| 58th | 1966 | Katahdin, ME | Caldwell |
| 59th | 1967 | Amherst, MA | Robinson, Drake, Foose |
| 60th | 1968 | New Haven, CT | Orville |
| 61st | 1969 | Albany, NY | Bird |
| 62nd | 1970 | Rangely Lakes, ME | Boone |
| 63rd | 1971 | Concord, NH | Lyons, Stewart |
| 64th | 1972 | Burlington, ME | Doolan, Stanley |
| 65th | 1973 | Frederick, NB | Grenier |
| 66th | 1974 | Orono, ME | Osberg |
| 67th | 1975 | Great Barrington, MA | Ratcliffe |
| 68th | 1976 | Boston, MA | Cameron |
| 69th | 1977 | Quebec City, PQ | Beland, LaSalle |
| 70th | 1978 | Calais, ME | Ludman |
| 71st | 1979 | Troy, NY | Friedman |
| 72nd | 1980 | Presque Isle, ME | Roy, Naylor |
| 73rd | 1981 | Kingston, RI | Boothroyd, Hermes |
| 74th | 1982 | Storrs, CT | Joeston, Quarrier |
| 75th | 1983 | Greenville - Millinocket, ME | Caldwell, Hanson |
| 76th | 1984 | Danvers, MA | Hanson |
| 77th | 1985 | New Haven, CT | Tracy |
| 78th | 1986 | Lewiston, ME | Newberg |
| 79th | 1987 | Montpelier, VT | Westerman |
| 80th | 1988 | Keene, NH | Bothner |

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# STRATIGRAPHY, STRUCTURAL GEOLOGY AND THERMOCHRONOLOGY OF THE NORTHERN BERKSHIRE MASSIF AND SOUTHERN GREEN MOUNTAINS

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## INTRODUCTION

The following text and generalized figures are intended to serve as the explanatory material for two fieldtrip guides, A-1 and B-1, contained in this volume. Trip A-1 covers an area at the northern end of the Berkshire massif in Massachusetts, and the southernmost end of the Green Mountains massif in Vermont and Massachusetts. Trip B-1 covers an area from Wilmington, Vt. in the Sadawga-Rayponda dome area, across the central Green Mountains to Bromley Mountain. Figures contained in the road log for Trip B-1 are numbered consecutively and follow those in the log for Trip A-1. References for both trips are cited at the end of the text in Trip A-1.

These two trips focus chiefly on the comparative structural geology and dynamothermal history of two major exposures of Middle Proterozoic basement rocks in the Northern Appalachians, as well as smaller exposures within the Sadawga-Rayponda domes and a complex zone of thrust faulted slivers of Middle Proterozoic rocks lying just east of the Green Mountains massif, south of Jamaica, Vt.

Stop locations and presentations are intended to present data related to the following general topics:

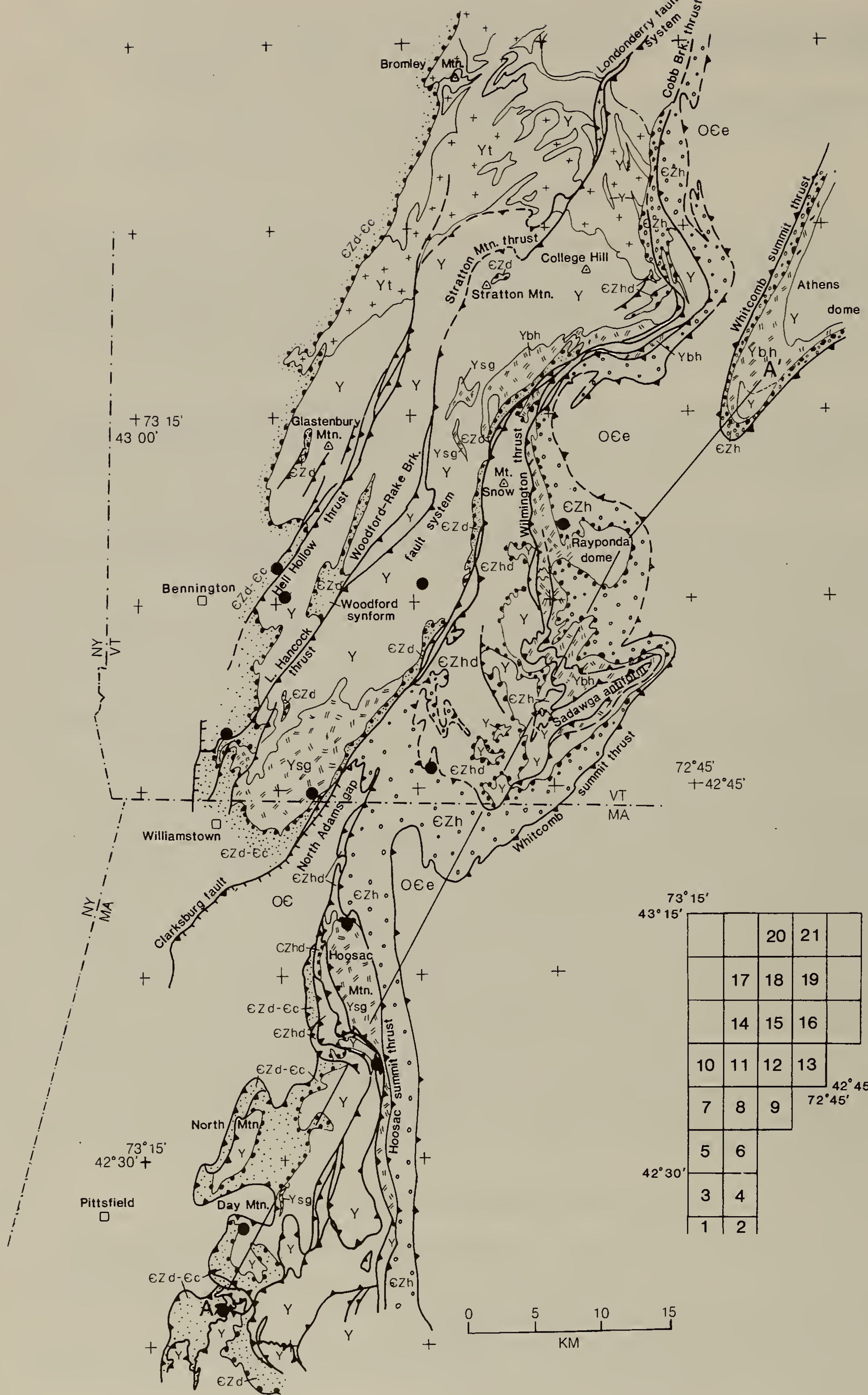
(1) Lithologic and chemical characteristics of the stratigraphic succession of the Middle Proterozoic rocks in each of the areas of basement rock;
(2) Characteristics of Proterozoic deformation;
(3) Stratigraphy and important facies relationships of the Late Proterozoic through Cambrian cover sequence rocks;
(4) Paleozoic deformation and dynamothermal effects of Taconic (about 460 Ma) and Acadian (about 380 Ma) orogenesis;
(5) The usefulness and applicability of 40Ar/39Ar mineral age spectra to unraveling the thermal history of complex polydeformed belts.

The results presented stem from detailed geologic mapping completed in the Berkshire massif and southern Green Mountains, before 1978 by N. Ratcliffe and S. Norton and recent 1:24,000 mapping by W. Burton and N. Ratcliffe in the area of the Green Mountains since 1986. Because the mapping is still in progress, much of the information in the Green Mountains is new and is preliminary. In the southern Green Mountains, pioneering studies by Skehan (1961) in the Wilmington, Vt., 15-minute quadrangle have been invaluable. His comprehensive work covered a large area of very complexly deformed, and locally, very poorly exposed rocks. We have adopted many of the basic elements of Skehan's work, but have added greater detail and definition in some areas. Recent published maps by Karabinos (1984) in the Jamaica area identify for the first time the location of potentially important faults along the eastern margin of the Green Mountains massif and Jamaica anticline, a zone that appears to extend southward into the Sadawga-Rayponda dome and along the eastern margin of the Green Mountains massif. U-Pb zircon studies conducted by Karabinos and Aleinikoff (1988) at the U.S. Geological Survey in Denver, on samples from basement rocks of the southern Green Mountains, and unpublished data of Samuel Mukasa and Beth Harding at University of Florida have helped resolve the age of the Stamford Granite Gneiss in Massachusetts, and of similar rocks in Vermont. 40Ar/39Ar data of Mukasa, from traverses across the Berkshire massif and southern Green Mountains (Sutter and others, 1985) have been supplemented by new data in the Green Mountains collected north of Mukasa's traverse in the area of trip B-1.

## LITHOTECTONIC UNITS AND MAJOR STRUCTURAL FEATURES

Major lithotectonic units of concern for these field trips within the area of figure 1 listed from lowest tectonic level to highest are:

(1) Green Mountain massif consisting of Middle Proterozoic gneiss and unconformable quartzofeldspathic cover, Dalton Formation and Cheshire Quartzite (CZd-Cc);
(2) Miogeoclinal rocks of the Stockbridge Valley exposed in the North Adams gap (CO);
(3) Berkshire massif allochthon consisting of nested thrust slices of Middle Proterozoic gneiss (Y) and its unconformable cover rocks of Dalton Formation (CZd);



Bromley Mtn.
Londonderry fault system
Cobb Brk. thrust
OЄe
Yt
Y
ЄZh
ЄZd-Єc
Stratton Mtn. thrust
ЄZd
College Hill
Stratton Mtn.
ЄZhd
Whitcomb summit thrust
Athens dome
Ybh
A'
Ysg
73 15'
43 00'
Glastenbury Mtn.
Mt. Snow
Wilmington thrust
Woodford-Rake Brk. fault system
Hell Hollow thrust
Rayponda dome
Bennington
Woodford synform
L. Hancock thrust
NY
VT
Sadawga antiform
72°45'
42°45'
VT
MA
Williamstown
North Adams gap
Whitcomb
Clarksburg fault
OЄ
Hoosac Mtn.
CZhd
Hoosac summit thrust
North Mtn.
73°15'
42°30'
Pittsfield
Day Mtn.
A
43°15'
20 21
17 18 19
14 15 16
10 11 12 13
7 8 9
5 6
3 4
1 2
0 5 10 15
KM

OЄ

Ordovician and Cambrian migeoclinal rocks

OЄe

Ordovician and Cambrian eugeoclinal rocks

ЄZd
ЄZd-Єc

Dalton Fm. or
Dalton and Cheshire
fms. combined

ЄZhd

Transitional
Hoosac-Dalton

ЄZh

Hoosac Fm.

Ysg
Ybh

**Ysg,** Stamford Granite Gneiss

**Ybh,** Bull Hill Gneiss of Doll and others, 1961

Y

Well-layered biotite
paragneiss, metavolcanics
and biotite granite gneiss

Yt

Biotite trondhjemitic gneiss,
metatonalite, metadacitic volcanics
and associated mafic metavolcanics

Contact

Unconformity, balls on younger rocks

Thrust fault, Taconic or Acadian

Normal fault, tick marks on downthrown side

field trip stop

Figure 1. Generalized, regional geologic map of the northern Berkshire massif, southern Green Mountains and Sadawga-Rayponda dome areas of Middle Proterozoic gneiss, showing fieldtrip stops for trips A-1 identified by solid dots. Geology is based on mapping by Ratcliffe and Burton since 1986 in the following quadrangles: 21-Londonderry, 20-Peru, 17-Sunderland, 18-Stratton Mtn., 19-Jamaica, 10-Pownal, 11-Stamford, 12-Readsboro, by Burton in 17-Sunderland, 14-Woodford, by Ratcliffe in 13-Jacksonville, and 15-Readsboro; Areas mapped previous to 1979, by Ratcliffe 1-East Lee, 3-Pittsfield East, 5-Cheshire, 7-Williamstown, 8-North Adams. Data in E. Dover (16), from Shehan, 1961; in Becket (2), Norton, 1974, in Windsor (6), Norton, 1961. Area just east of Jamaica (19) from Karabinos, 1984. (For more detailed map of the Green Mountains see figure 15, trip B-1).

(4A) Hoosac Nappe and its unconformable Hoosac Formation cover (CZhd) and its lateral equivalent to the north;
(4B) Basement rocks of the Sadawga-Rayponda dome below the Wilmington fault system (Ybg) and other Y units);
(5) Allochthonous Hoosac cover (CZh) and Bull Hill Gneiss of Doll and others (1961), (Ybh) above the Wilmington fault and Hoosac summit thrust;
(6) Allochthonous Rowe Schist and eugeoclinal rocks (OCe) above the Whitcomb summit thrust. The Whitcomb summit thrust marks the westernmost occurrence of eugeoclinal rocks and the present approximate position of the root zone for the Taconic allochthons according to Stanley and Ratcliffe (1985).

The boundaries between the major lithotectonic units are ductile thrust faults of probable Taconian age. All of the major faults are folded, and locally form recumbent folds. A characteristic, mylonitic fabric parallels most boundaries. This mylonitic fabric is expressed by a strong rodding and foliation that commonly contains oriented minerals in an assemblage compatible with the highest grade of Paleozoic metamorphism in an area. In much of the area shown in figure 1 post-thrust mineral assemblages of probable Acadian age generally overgrow fault fabrics (Stop A-3). On the east flank of the Sadawga-Rayponda dome, extending northward to a point east of Jamaica, particularly coarse-grained static overgrowth rims on garnets, randomly oriented sprays of amphibole, and abundant late plagioclase, microcline, and quartz feldspar pegmatitic pods all of which lack mylonitic fabric are present in the fault zones.

Within the core of the Green Mountains massif, several fault zones (fig. 1) have been mapped. These faults are accompanied by wide zones of extensive ductile deformation and pronounced zones of dynamothermal retrogression. Chlorite to biotite-grade phyllonites and sericite-rich mylonite gneiss are common, in zones containing highly complex folds, abundant sheath folds and lineated tectonites. These widely spaced zones of accumulated strain appear to the principal mechanism for deformation within the Green Mountain massif. Three such zones, the Hell Hollow thrust, Lake Hancock-Woodford-Rake Brook fault zones, and the Londonderry-Stratton Mountain faults strike northeast and dip moderately to steeply southeast. These faults and ductile deformation zones extend diagonally from northeast to southwest across the massif (fig. 1). The Lake Hancock-Woodford-Rake Brook system of faults mark a well-defined deformational front, east of which Middle Proterozoic rocks contain a penetrative Paleozoic foliation of biotite grade or higher. Based on limited biotite 40Ar/39Ar data, the penetrative fabric associated with this fault zone and east of it is Taconian (Sutter and others, 1985) (see discussions for stops 9 and 10 of trip A-1 and stops 5, 9, 10 and 11 of trip B-1). In the western part of the massif, Paleozoic fabric is less penetrative, and is more restricted to fault zones.

The Searsburg fault system forms a continuous thrust fault along the eastern margin of the Green Mountains massif from Jamaica south to the Massachusetts state line where it either merges with or is offset by the Clarksburg fault system. Brittle faults and associated open-work breccias, hematite-coated joint surfaces and sulfide mineralization mark certain exposures of the gently southeast-dipping Clarksburg fault from a point east of Williamstown north to near Stamford, Vermont, and late, low-angle Mesozoic(?) extensional faulting is likely along this zone (fig. 1). If this interpretation is correct, the older Searsburg thrust faults may correlate with the Hoosac fault and nested thrust faulting on the lower surface of the Hoosac slice seen at stop 3, trip A-1.

A cross section drawn from the beginning of Trip A-1 in the Berkshire massif through the Sadawga and Rayponda domes to the Athens dome, shows diagrammatically, the correlation of lithotectonic units and major faults (fig. 2). Projected positions of fieldtrip stops for trip A-1 are shown. The Hoosac summit thrust, although well-defined on Hoosac Mountain (A-1, stop 3), cannot be correlated with complete confidence. In figures 1 and 2, the Hoosac summit thrust is connected along a highly contorted contact within the Hoosac Formation with the Wilmington fault system. The Wilmington fault system is continuous with the Cobb Brook thrust of Karabinos (1984). This zone of fault-intercalated rocks may reappear within the Chester-Athens dome as the inner contact between Bull Hill Gneiss and Hoosac Formation (shown as the Cavendish Formation in Doll and others, 1961).

The Whitcomb summit thrust is drawn through areas we have not mapped and both its location and its existence as a discrete fault is in part conjectural. In areas where we have mapped it, in the North Adams and Jamaica quadrangles, both physical evidence for faulting (mylonitic structures) and truncation of map units supports a fault within this part of the section.

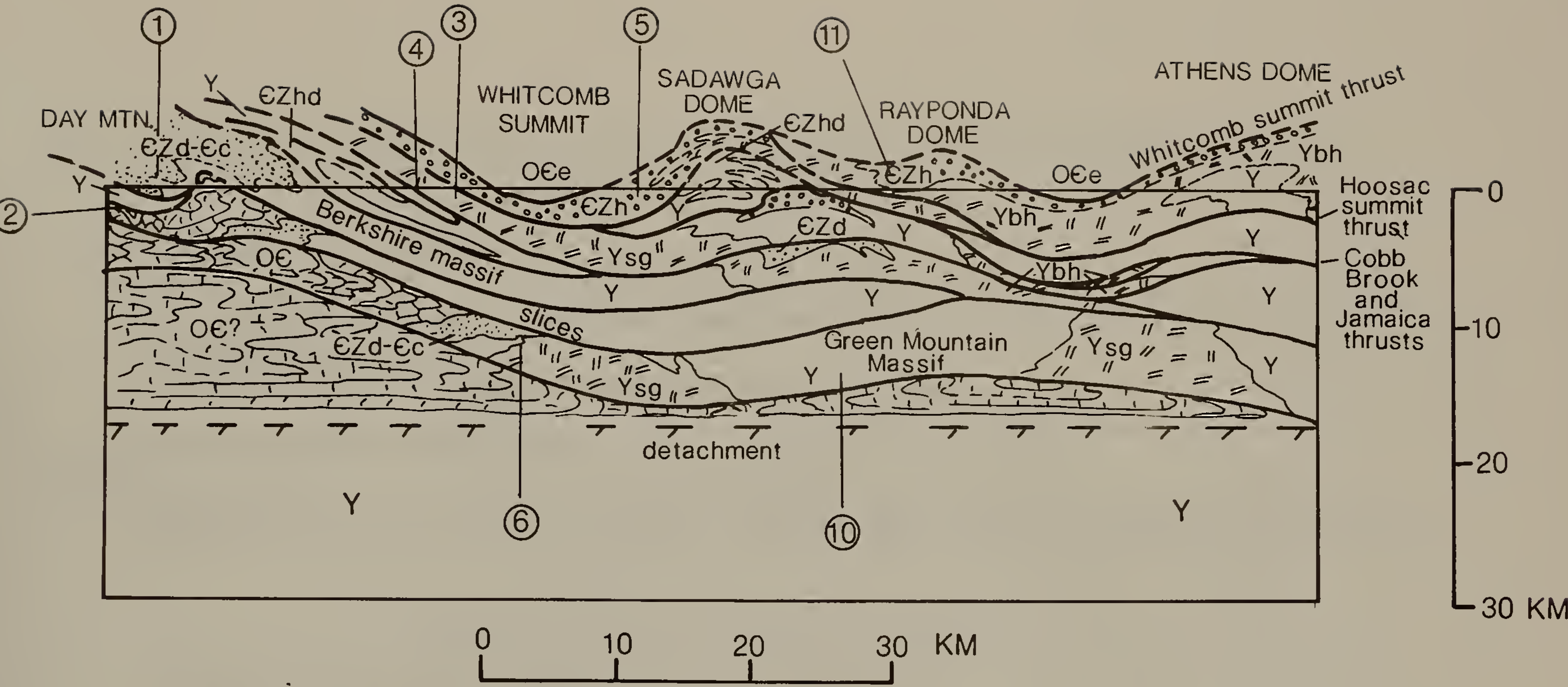



Figure 2. Generalized cross section from Day Mountain northeastward to the Athens dome showing correlation of faults and lithotectonic units, numbered localities refer to projected approximate position of stops on Trip A-1. For similar diagram pertinent to Trip B-1 see figure 17 in that trip log.

## STRATIGRAPHY

### Middle Proterozoic rocks

Hornblende granulite metasedimentary, metavolcanic, and syn-deformational granitoids greater than 1 Ga form the bulk of the parautochthonous Middle Proterozoic basement rocks of the Berkshire and Green Mountain massifs.

Middle Proterozoic rocks of the Green Mountains and Berkshire massifs may be grouped in 3 major categories based on field observations and U-Pb zircon ages. These are: (1) post-tectonic granites and pegmatite; (2) syntectonic granitoids and migmatitic granitic gneisses; and (3) a pretectonic layered paragneiss and metavolcanic sequence (see table 1). In addition, a newly recognized pretectonic suite of metatrondhjemite, metatonalite, metadacite and hornblende gneisses (meta-andesite and basalt) shown in table 2 is of uncertain stratigraphic position at present. It crops out in the central Green Mountains in the area of trip B-1, north of Stratton Mountain.

Mapping in the southern Green Mountains has confirmed that the sequence of Middle Proterozoic gneisses present in the Berkshire massif (Zen, 1983) is also present within the Green Mountains. Likewise, the basement gneiss within the Sadawga-Rayponda dome and in the small areas of gneiss between Mount Snow and Jamaica correlate with rocks present in the Green Mountains massif to the west. This stratigraphic and plutonic succession (table 1) is present in the Green Mountains as far north as Stratton Mountain. North of that point a belt of distinctive trondhjemitic and tonalitic gneisses approximately 10 kilometers wide as measured in a north- south direction extends across the full width of the massif (table 2).

The Stamford Granite Gneiss consists of coarse-grained to pegmatitic, microcline- microperthite-megacrystic, rapakivi granite; medium-grained, aplitic granite; hornblende-biotite monzonite, and locally ferromonzonitic mafic segregations and dikes. The rock clearly post-dates Grenville tectonism as it crosscuts folds in the gneisses and lacks the gneissic structure of the country rocks. The unit is present in the southern Green Mountains, on Hoosac Mountain and in the Green Mountains south of Jamaica from Wardsboro southeast to a point just west of Mount Snow. In the latter exposures, K-feldspar megacrystic biotite granite is largely transformed by Paleozoic shearing to biotite-augen gneiss. Mylonite and mylonite gneiss are co-extensive with the Bull Hill Gneiss as mapped by Doll and others (1961) near Wardsboro, Vt. and in the Sadawga- Rayponda dome. A concordant U-Pb zircon age of 959?Ma from the type Stamford Granite Gneiss (stop 5, trip A-1), (Karabinos and Aleinikoff, 1988) and similar but slightly more discordant ages by them from the Stamford-like rocks in the belt near Wardsboro and from the Bull Hill gneiss of Doll and others (1961) from the Chester and Athens domes indicate that all these granites are coeval. A U-Pb, concordia zircon age of approximately 960 Ma has also been determined by Beth Harding and Samuel Mukasa from the Stamford Granite Gniess on Hoosac Mountain. The Stamford Granite Gneiss

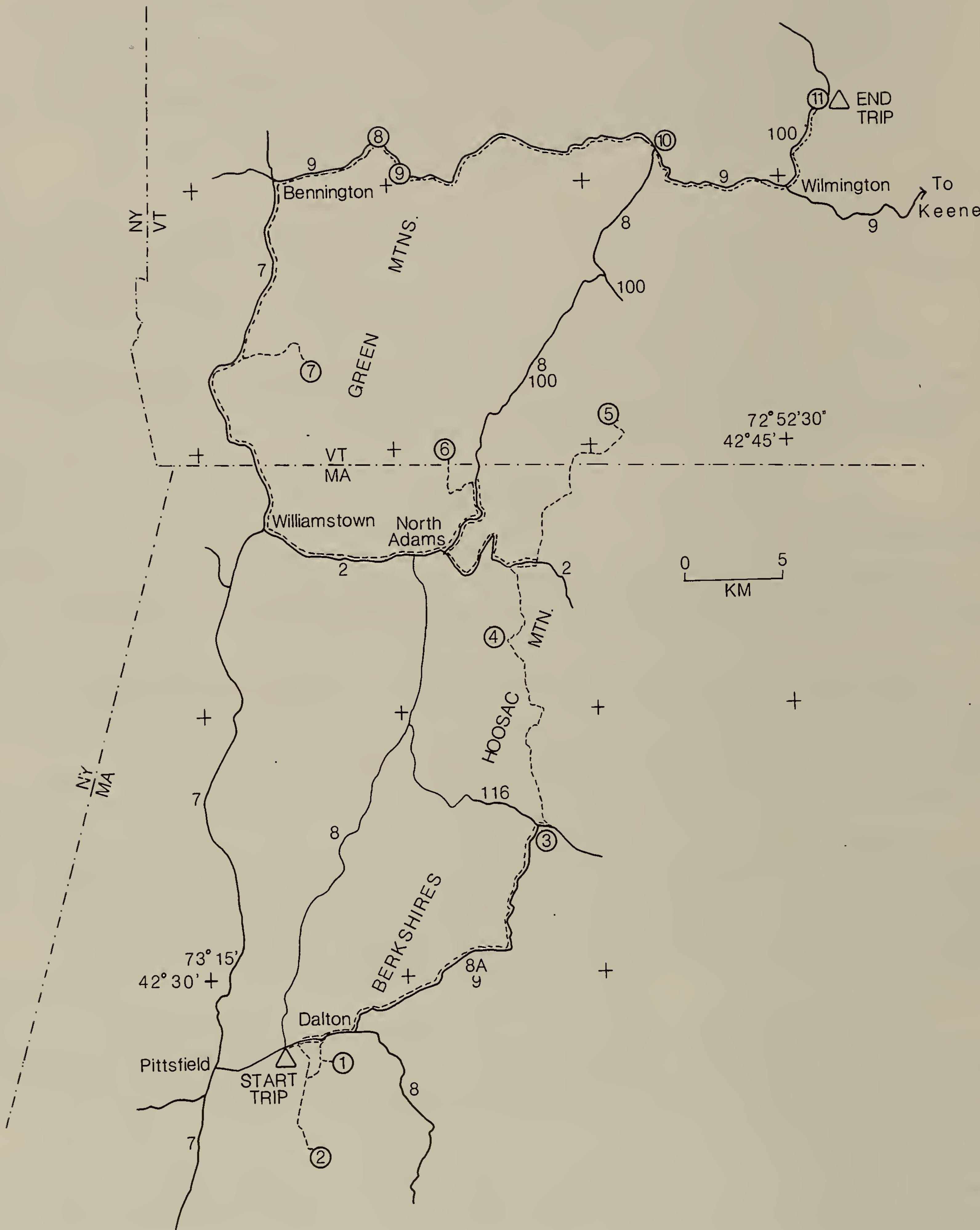



Figure 3. Map showing location of A-1 field trip.

Table 1. Generalized stratigraphic-plutonic succession for Middle Proterozoic rocks of the Berkshire massif and southern Green Mountains area (units shown on figures 5 and 6)

A. Post-tectonic granite and pegmatite

Stamford Granite Gneiss (Ysg)

Coarse-grained to pegmatite, microcline-microperthite megacrystic, biotite-rapakivi granite, associated ferromonzonite and aplite
Correlative with the Bull Hill Gneiss (Ybh) of Doll and others (1961)

B. Syntectonic biotite or hornblende granites

Tyringham Gneiss (Ytg)

Coarse-grained, microcline-perthite megacrystic, hastingsite-biotite granodiorite and granite correlative with other biotite-Kspar megacrystic granite gneiss shown Ygg by Zen (1983) in the Berkshire massif

Biotite granite gneiss of the Green Mountains (Ygg)

Medium-grained, well-foliated, biotite Kspar-rich, locally megacrystic, granitic gneiss exposed on Glastenbury Mountain and in the Stratton Mountain area (Ygg)

Biotite, migmatitic-K-feldspar-plagioclase gneiss (Ygm)

Medium-grained, biotite-plagioclase-Kspar-quartz, gneiss having ghost-like inclusions of well-layered 2 feldspar biotite gneiss

Biotite-K-feldspar megacrystic granite gneiss on College Hill (Ygc)

Exposed on a belt extending from College Hill in the Jamaica, Vt., quadrangle to a point west of Stratton Mountain

C. Layered paragneiss and metavolcanics gneisses (not necessarily listed in stratigraphic order)

Biotite-quartz-plagioclase gneiss (Ybg)

Well-layered, white and black, biotite-quartz-plagioclase gneiss and thin interlayered amphibolite, includes amphibolite, hornblende-pyroxene calcsilicate rocks, diopside- plagioclase gneiss, and rare beds of calcite-diopside marble

Quartzite (Yq)

White to steel-gray, vitreous, garnet-biotite-quartzite in beds up to 5 m thick, associated with garnet-rich, biotite-quartz plagioclase gneiss

Felsic gneiss (Yfg)

White-weathering, very fine-grained, magnetite-biotite-quartz-plagioclase-microcline gneiss, metarhyolite

Rusty-weathering, garnet-sillimanite, quartz gneiss (also shown as Yq)

Commonly retrograded to chlorite-muscovite-quartz phyllite that locally contains chloritoid in the Green Mountains

Amphibolite and amphibolite gneiss (Ya)

(In the Green Mountains)

Lee Gneiss (in the Berkshire massif)

Dark-colored, hornblende, biotite-plagioclase gneiss, amphibolite and dioritic gneiss (probable intermediate to mafic volcanics)

Washington Gneiss (Yw), (Yrr)

Rusty weathering, garnet-biotite-plagioclase quartz gneiss and schist, marked by coarse ribs of lavender quartz, (Yrr in the Green Mountains) associated with graphite-rich quartzite, metaconglomerate and sulfidic graphitic marble, and in Massachusetts, including a biotite quartz-plagioclase leucogneiss interpreted as metadacite and garnet amphibolite interpreted as mafic volcanic rocks

Table 2. Generalized sequence of rocks in the metatonalite-metatrondhjemite gneiss belt, central Green Mountains massif (Peru, Londonderry quadrangles)

| | |
|---|---|
| Ytr | White weathering, massive, medium- to coarse-grained biotite trondhjemitic gneiss (well exposed near Rawsonville and on the crest of Bromley Mtn) characterized by 0.5 to 1 cm rectangular non-oriented clots of biotite and well-twinned subhedral grains of oligoclase |
| Ytm | White weathering, medium- to fine-grained, biotite-quartz-plagioclase granofels having indistinct layers accentuated by rare thin layers of pegmatitic granodiorite, giving rock a migmatitic appearance |
| Ytd | White weathering, very fine-grained, quartz plagioclase rock, having less than 5 percent biotite, and indistinct wispy biotite-rich laminae, locally, well-layered having biotite-rich, quartz-plagioclase interlayers up to 0.25 m thick, probably metadacite, well exposed at Bondville |
| Yt | Massive, coarse-grained, hornblende-biotite-quartz plagioclase metatonalite, characterized by 10-20% mafic minerals, distinct relict igneous textures, and zones rich in dioritic (cognate?) xenoliths, well-exposed at Cole Pond in the Londonderry quadrangle |
| Yta | Black, hornblende amphibolite and black and white equigranular hornblende plagioclase gneiss, locally very well-layered on a 0.5 m to 1 m scale, locally abundant veins of hornblende or biotite diorite or tonalite, well-exposed at South Londonderry |

Table 3. Representative chemical analyses of Stamford Granite Gneiss from the Green Mountain and from Hoosac Mountain Analyses by rapid rock techniques described in Shapiro (1975). N. Rait, H. Smith analysts.

| | --------------Stamford, VT | | | | ---------- | ------Hoosac Mtn.----- | |
|---|---|---|---|---|---|---|---|
| Sample No. | 170 | 171 | 173 | 6 | 172 | 102-32 | 176 |
| | --------rapakivi granite | ------- | | border | dike | | |
| Constituent wtg% | | | | | | | |
| $SiO_2$ | 67.6 | 69.2 | 70.7 | 64.5 | 50.5 | 68.0 | 69.4 |
| $Al_2O$ | 16.0 | 16.4 | 14.9 | 14.8 | 14.9 | 15.4 | 14.0 |
| $Fe_2O$ | 1.5 | 1.1 | 0.90 | 1.8 | 2.6 | 2.0 | 1.9 |
| Feo | 1.7 | 1.2 | 1.7 | 4.8 | 8.5 | 2.2 | 2.6 |
| MgO | 0.17 | 0.12 | 0.20 | 1.1 | 3.5 | 0.36 | 0.82 |
| CaO | 2.4 | 2.2 | 2.1 | 3.1 | 7.7 | 2.0 | 0.83 |
| $Na_2$ | 3.5 | 3.5 | 3.2 | 3.2 | 2.2 | 3.6 | 2.8 |
| $K_2O$ | 5.9 | 6.4 | 5.8 | 4.2 | 3.7 | 5.4 | 5.8 |
| $H_2O^+$ | 0.55 | 0.32 | 0.37 | 0.62 | 1.0 | 0.16 | 0.04 |
| $H_2O^-$ | 0.02 | 0.02 | 0.08 | 0.03 | 0.06 | 0.04 | 0.03 |
| $TiO_2$ | 0.29 | 0.23 | 0.24 | 1.1 | 2.5 | 0.39 | 0.42 |
| $P_2O_5$ | 0.15 | 0.14 | 0.14 | 0.45 | 1.0 | 0.19 | 0.18 |
| MnO | 0.04 | 0.03 | 0.04 | 0.07 | 0.15 | 0.06 | 0.04 |
| $CO_2$ | 0.01 | 0.08 | 0.03 | 0.07 | 1.5 | 0.01 | 0.03 |
| Totals% | 99.83 | 100.94 | 100.40 | 99.94 | 98.9 | 99.86 | 98.89 |
| Total S% | 0.12 | 0.09 | 0.077 | 0.093 | 0.97 | 0.027 | 0.14 |
| CIPW NORMATIVE MINERALS | | | | | | | |
| apatite | 0.36 | 0.34 | 0.34 | 1.08 | 0.43 | 0.44 | 0.44 |
| ilmenite | 0.56 | 0.44 | 0.45 | 2.09 | 0.74 | 0.80 | 0.65 |
| orthoclase | 35.0 | 37.40 | 34.6 | 25.05 | 31.9 | 34.5 | 35.2 |
| albite | 29.9 | 29.4 | 27.3 | 27.06 | 30.5 | 23.9 | 26.5 |
| anorthite | 10.5 | 10.3 | 9.1 | 13.8 | 9.8 | 3.87 | 6.40 |
| corundum | 0.0 | 0.0 | 0.0 | 0.0 | 0.03 | 1.72 | 0.9 |
| magnetite | 2.2 | 1.59 | 1.02 | 2.64 | 2.9 | 2.75 | 3.08 |
| diopside | 1.14 | 0.43 | 0.0 | 1.11 | 0.0 | 0.0 | 0.0 |
| hypersthene | 1.80 | 1.52 | 2.8 | 7.26 | 2.70 | 4.6 | 0.86 |
| quartz | 18.43 | 18.58 | 24.3 | 20.0 | 20.78 | 27.10 | 25.61 |
| Total | 99.86 | 100. | 99.91 | 100.09 | 99.78 | 99.68 | 99.64 |

on Hoosac Mountain, the Bull Hill Gneiss (of Doll and others, 1961) in the Chester and Athens domes and similar K-feldspar megacrystic granite gneiss in the Jamaica-Wardsboro area were originally believed to be metarhyolitic volcanic rocks interlayered with various cover sequence rocks, either the Hoosac Formation (Norton, 1969) or the Cavendish Formation (of Doll and others, 1961). The U-Pb data, coupled with the observations that: (1) the Stamford and Stamford-like rocks intrude Grenvillian basement, (2) they are unconformably overlain by the Hoosac Formation and (3) they form fault slivers within the cover sequence (fig. 1) suggest that rocks previously mapped as the Bull Hill Gneiss of Doll and others (1961) in the Chester and Athens domes may locally also be faulted into the cover sequence.

These very distinctive, coarse-grained, 960-Ma-old granites and pegmatite granites are unique in the Appalachians to the basement rocks of the southern Green Mountains, northern Berkshire massif, Sadawga-Rayponda, and Chester and Athens domes. Chemical data suggest that, although the granites are all very similar K-rich granites (fig. 4) each individual area is slightly different (see discussion stops A-3, 6). The field data suggest that many relatively small, separate rapakivi plutons formed in a general east-west belt across the Grenville-deformed basement rocks. Chemistry and structural setting suggest that these may mark a post-Grenville anorogenic event, much older than the start of the Late Proterozoic rifting event responsible for the formation of Iapetus.

The occurrence of these rocks within these areas indicates that until 960 Ma ago, the Proterozoic rocks of the Green Mountains-Chester Athens domes, Sadawga-Rayponda dome and Berkshire massif were all part of the pre-Iapetan, or Laurentian continent.

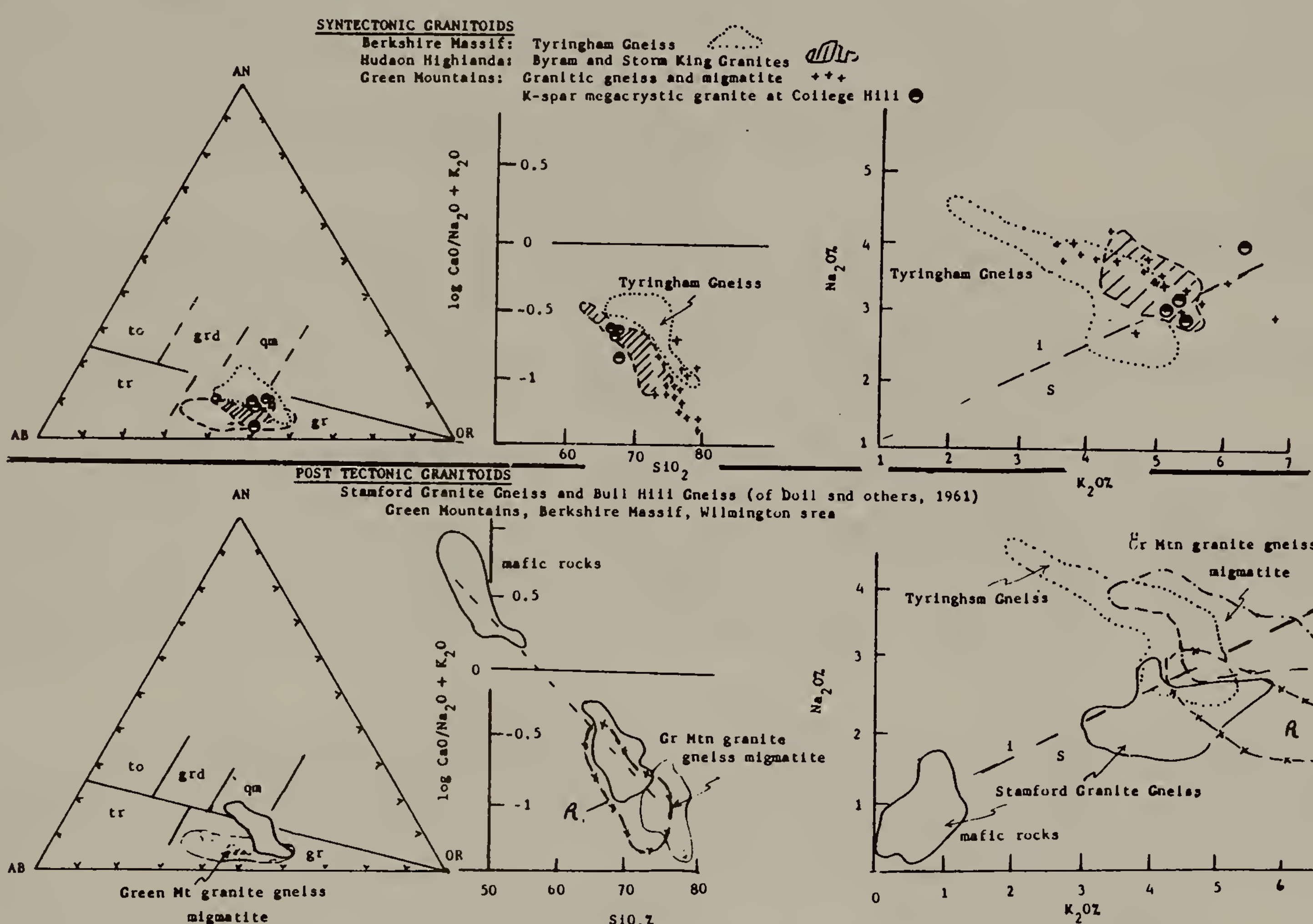



Figure 4. Selected chemical characteristics of syntectonic and post-tectonic granitoids in Green Mountains and Berkshire massifs. AN, AB, and OR - normative anorthite, albite, orthoclase diagram showing classification scheme of O'Connor, 1965; to = tonalite, tr = trondhjemite, grd = granodiorite, qm = quartz monzonite, gr = granite; I, S in $Na_2O$ vs. $K_2O$ plot refers to igneous and sedimentary granite fields of Chappel and White, 1974. Shaded fields identify granitic gneiss and migmatites from Green Mountains in relation to other rocks. Sources of data: Bryam and Storm King Granites (Drake, 1984; Lowe, 1950; Helenek and Mose, 1984); other data (Ratcliffe, unpub. data). R and dash-x'd field rapakivi granites of Finland from Nurmi and Haapala, 1986.

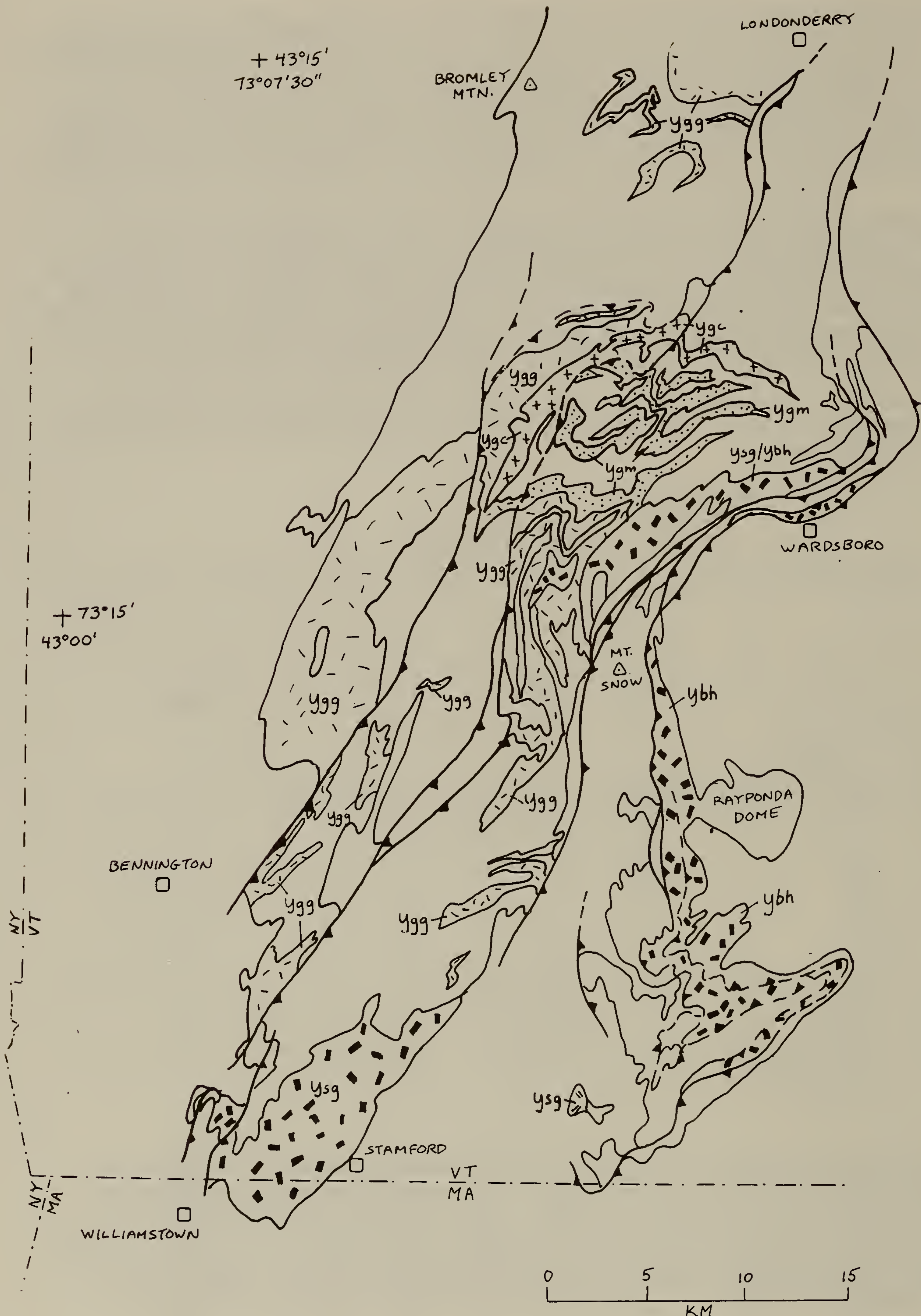



Figure 5. Generalized geologic map showing distribution of syntectonic and post-tectonic Middle Proterozoic granitoids in the Green Mountains and Sadagwa-Rayponda domes identified by symbol is table 1.

Granitic gneisses interpreted as syntectonic intrusives form approximately one third of the area of the Berkshire massif (the Tyringham and other granitic gneisses). These units are in contact with all paragneiss units and exhibit locally clear, crosscutting relationships across gneissic layering. Chemically similar crosscutting but highly deformed granitoids are common in the Hudson Highlands and Reading Prong (the Storm King Granite of Lowe, 1950) and the Byram Intrusive Suite of Drake (1984). Similar biotite granitoids in the Green Mountains (table 1) are also widely distributed (fig. 5). Chemical analyses (tables 4 and 5) show that the Green Mountain granitoids are all potassic granites (fig. 4) having restricted $CaO/Na_2O+K_2O$ vs $SiO_2$ ranges and calc-alkaline trends, similar to syntectonic granites in the Hudson Highlands and Berkshire massif. Migmatitic granite gneisses (Ygm) are all very $SiO_2$-rich and range greatly in $CaO/Na_2O+K_2O$ values, perhaps reflecting some sedimentary component in the local source rocks (table 5). These migmatitic gneisses do not show crosscutting relations to country rocks.

Paragneiss and metavolcanic rocks

A distinctive group of metasedimentary gneisses and metavolcanic rocks (rhyolites, andesites, and basaltic rocks) (table 1 and figure 6), are present in the Berkshire massif and Green Mountain massif as far north as Stratton Mountain. Distribution of selected units identified in table 1, are shown in figure 6. The bulk of the section (Ybg) is well-layered biotite plagioclase gneiss that forms the host for zones of calc-silicate rocks, marbles, and amphibolites, and garnet-rich-biotite-plagioclase gneiss. Locally, quartzites up to 10 m thick are present (Yq). Sillimanite-garnet gneiss and schist is present near the quartzites, as are abundant white, K-feldspar-rich pegmatites. Garnet-quartzites are typically mylonitic, having a high-grade ductile fabric in which broken, almond-shaped fragments of orangish red garnet float in a mylonitic matrix of quartz and sillimanite. Where intruded by pegmatite, sillimanite is retrograded to muscovite, and subsequent Paleozoic dynamothermal retrogression has produced a fine-grained chlorite-muscovite-chloritoid phyllite or phyllonite from the Proterozoic rocks. Exposures of rocks like this cap the peaks of Bromley Mountain, Stratton Mountain, and the peaks west of College Hill. These rocks could be easily mistaken for Paleozoic aluminous cover rocks such as the Gassetts Schist member of the Cavendish Formation of Doll and others, 1961, or aluminous rocks in the Hoosac Formation.

Rusty, sulfidic, sillimanite-biotite-quartz-ribbed gneiss, known as the Washington Gneiss in Massachusetts and Connecticut, is interpreted as the base of the section and is present as far north as the Mount Snow quadrangle. It is shown as Yrr in Figure 6. This unit appears to thin northward from Massachusetts.

Pretectonic metavolcanic rocks from the Berkshire massif, contained within the Ybg unit, the Lee Gneiss, and a felsic metarhyolite unit define a calc-alkaline suite having moderate iron enrichment (fig. 7).

The metatonalite-metatrondhjemite sequence (Yt on figs. 1 and 6) is only present within the basement rocks of the Green Mountains. It is characterized by equal abundance of three major rock types: (1) white weathering, very fine-grained, quartz plagioclase gneiss having rare thin biotite-rich layers 1 cm to 0.5 m thick--probably a metadacite; (2) white weathering, coarse- grained, biotite spotted metatrondhjemite, characterized by rectangular spots of biotite 0.5 to 1 cm, possibly pseudomorphic after pyroxene; and (3) coarse-grained, non-layered, hornblende- biotite metatonalite and associated hornblendic gneisses and amphibolite, (probably intrusive tonalite and mafic metavolcanics).

Within this suite (table 2), the coarse-grained, more leucocratic gneisses intrude the more mafic members. Irregular inclusions of mafic dioritic gneiss are common throughout, and, at some localities, angular xenoliths exist. Metatonalite commonly contains irregular to subrounded inclusions of what appear to be hornblende-biotite diorite as cognate xenoliths. Over broad areas, a distinctive well-layered, biotite gneiss, calcsilicate and aluminous quartzite sequence parallels the contact with the adjacent metatonalite/metatrondhjemite suite, thus suggesting that the paragneiss sequence may unconformably overlie the Yt unit.

Limited chemical data, table 6 and figure 5, indicate the rocks of the Yt suite are closely related, as they define a tight linear array on log $CaO/Na_2O+K_2O$ and AFM diagrams (fig. 7), indicative of a calc-alkaline to calcic plutonic suite. In normative An-Ab-Or classification, these rocks are tonalites and trondhjemites (or dacites and keratophyres).

The suite has the same very high $Na_2O/K_2O$ values and overall chemistry as members of the Losee Metamorphic Suite and related rocks of the Reading Prong-Hudson Highlands of New York, New Jersey, and Pennsylvania (fig. 7), generally held to be dacitic and quartz keratophyric volcanic rocks (Drake, 1969).


Figure 6. Generalized geologic map showing distribution of pretectonic Middle Proterozoic paragneiss and metavolcanic units in the Green Mountains and Sadagwa-Rayponda domes identified by symbol in table 1.

Table 4. Chemical analyses of biotite-plagioclase-microcline granitic gneiss at College Hill, Peru quadrangle, VT (Ygc) and biotite granite gneiss (Ygg)

| Sample | 3348A | 3348B | 3348C-1 | 3348C-2 | 849 | 851 | 444 | 910 | 930 | 241 |
|---|---|---|---|---|---|---|---|---|---|---|
| | Ygc | | | | Ygg | | | | | |
| **constituent** | | | | | | | | | | |
| $SiO_2$ | 69.9 | 68.9 | 67.8 | 68.5 | 73.70 | 76.10 | 78.30 | 72.70 | 71.60 | 73.30 |
| $Al_2O_3$ | 15.2 | 13.7 | 13.3 | 13.5 | 12.70 | 12.90 | 10.80 | 14.20 | 14.20 | 14.00 |
| FeO | 0.66 | 1.42 | 1.36 | 1.40 | 1.50 | 0.45 | 0.99 | 0.44 | 1.20 | 0.21 |
| $Fe_2O_3$ | 1.2 | 2.9 | 4.0 | 3.5 | 0.92 | 0.23 | 0.53 | 0.98 | 0.82 | 0.54 |
| MgO | 0.53 | 0.56 | 0.77 | 0.69 | 0.28 | 0.33 | 0.15 | 0.34 | 0.53 | 0.17 |
| CaO | 0.69 | 1.81 | 2.17 | 1.91 | 1.02 | 0.46 | 0.78 | 1.26 | 1.38 | 0.75 |
| $Na_2O$ | 3.97 | 3.22 | 3.12 | 2.88 | 3.12 | 3.14 | 2.56 | 3.33 | 3.23 | 2.83 |
| $K_2O$ | 6.38 | 5.34 | 4.91 | 5.51 | 5.20 | 5.71 | 4.65 | 5.33 | 5.47 | 6.65 |
| $Tio_2$ | 0.21 | 0.59 | 0.80 | 0.72 | 0.31 | 0.05 | 0.17 | 0.22 | 0.31 | 0.07 |
| $P_2O_5$ | 0.07 | 0.19 | 0.29 | 0.23 | 0.07 | 0.05 | 0.05 | 0.07 | 0.12 | 0.05 |
| MnO | <0.02 | 0.06 | 0.08 | 0.07 | 0.03 | 0.02 | 0.02 | 0.02 | 0.02 | 0.02 |
| $H_2O^+$ | 0.30 | 0.35 | 0.45 | 0.53 | 0.40 | 0.33 | 0.30 | 0.27 | 0.33 | 0.51 |
| $H_2O^-$ | 0.06 | 0.25 | 0.31 | 0.24 | 0.11 | 0.11 | 0.22 | 0.18 | 0.18 | 0.16 |
| $CO_2$ | 0.01 | 0.01 | 0.02 | 0.03 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 |
| Total | 99.20 | 99.30 | 99.38 | 99.71 | 99.36 | 99.98 | 99.52 | 99.34 | 99.39 | 99.26 |
| **CIPW Norms** | | | | | | | | | | |
| **constituent** | | | | | | | | | | |
| Q | 20.0 | 24.6 | 24.2 | 24.7 | 32.9 | 34.6 | 43.7 | 30.4 | 28.3 | 30.1 |
| C | 0.7 | 0.0 | 0.0 | 0.0 | 0.3 | 0.8 | 0.3 | 0.9 | 0.8 | 0.9 |
| or | 38.2 | 32.0 | 29.5 | 33.0 | 30.1 | 33.9 | 27.8 | 31.9 | 32.7 | 39.8 |
| ab | 34.0 | 27.7 | 26.8 | 24.7 | 26.7 | 26.8 | 21.9 | 28.5 | 27.6 | 24.3 |
| an | 2.9 | 7.3 | 7.9 | 7.7 | 4.6 | 2.0 | 3.5 | 5.7 | 6.0 | 3.4 |
| di | 0.0 | 0.4 | 0.8 | 0.1 | | | 0.0 | 0.0 | 0.0 | 0.0 |
| hy | 2.7 | 4.5 | 6.7 | 5.9 | 2.2 | 1.4 | 1.5 | 0.9 | 2.4 | 0.4 |
| mt | 1.0 | 2.1 | 2.0 | 2.1 | 1.3 | 0.3 | 0.8 | 0.9 | 1.2 | 0.5 |
| il | 0.4 | 1.1 | 1.5 | 1.4 | 0.6 | 0.1 | 0.3 | 0.4 | 0.6 | 0.1 |
| hem | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.4 | 0.0 | 0.2 |
| ap | 0.2 | 0.4 | 0.7 | 0.5 | 0.2 | 0.1 | 0.1 | 0.2 | 0.3 | 0.1 |
| cc | 0.0 | 0.0 | 0.1 | 0.1 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 |
| total | 100.1 | 100.1 | 100.2 | 100.2 | 99.9 | 100.0 | 99.9 | 99.8 | 99.9 | 99.8 |

Table 5. Chemical analyses of migmatitic biotite feldspar gneiss (Ygm), Jamaica and Stratton Mountain quadrangles, VT

| Sample | 2831 | 1060A | 1060B | 1061 | 1058A | 675 |
|---|---|---|---|---|---|---|
| **constituent** | | | | | | |
| $SiO_2$ | 73.3 | 75.9 | 75.80 | 76.20 | 76.30 | 71.00 |
| $Al_2O_3$ | 13.0 | 12.10 | 12.10 | 12.20 | 13.50 | 13.80 |
| FeO | 1.2 | 1.15 | 1.49 | 0.57 | 0.28 | 1.40 |
| $Fe_2O_3$ | 1.2 | 0.81 | 0.66 | 1.10 | 0.43 | 2.30 |
| MgO | 0.35 | 0.21 | 0.10 | 0.18 | 0.10 | 0.11 |
| CaO | 0.77 | 0.47 | 0.35 | 0.78 | 0.83 | 0.63 |
| $Na_2O$ | 3.02 | 3.45 | 3.76 | 3.58 | 3.94 | 3.43 |
| $K_2O$ | 5.56 | 5.11 | 4.84 | 4.36 | 3.78 | 6.02 |
| $TiO_2$ | 0.28 | 0.19 | 0.21 | 0.16 | 0.02 | 0.26 |
| $P_2O_5$ | 0.06 | 0.05 | 0.05 | 0.05 | 0.18 | 0.07 |
| MnO | 0.02 | 0.02 | 0.02 | 0.02 | 0.08 | 1.20 |
| $H_2O^+$ | 0.11 | 0.31 | 0.25 | 0.48 | 0.27 | 0.33 |
| $H_2O^-$ | 0.27 | 0.10 | 0.11 | 0.10 | 0.20 | 0.12 |
| $Co_2$ | 0.01 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 |
| Total | 99.15 | 99.87 | 99.74 | 99.78 | 99.91 | 99.49 |
| **CIPW Norms** | | | | | | |
| **constituent** | | | | | | |
| Q | 32.6 | 34.7 | 34.2 | 36.6 | 37.4 | 26.9 |
| C | 0.8 | 0.2 | 0.2 | 0.3 | 1.9 | 0.8 |
| or | 33.3 | 30.4 | 28.8 | 26.0 | 22.5 | 35.9 |
| ab | 25.9 | 29.4 | 32.1 | 30.6 | 33.5 | 29.3 |
| an | 3.4 | 2.0 | 1.4 | 3.6 | 3.0 | 2.3 |
| hy | 1.7 | 2.0 | 1.8 | 1.7 | 1.0 | 0.6 |
| mt | 1.8 | 1.0 | 1.1 | 0.9 | 0.4 | 3.3 |
| il | 0.5 | 0.4 | 0.4 | 0.3 | 0.0 | 0.5 |
| ap | 0.1 | 0.1 | 0.1 | 0.1 | 0.4 | 0.2 |
| Total | 100.1 | 100.2 | 100.1 | 100.1 | 100.1 | 99.8 |

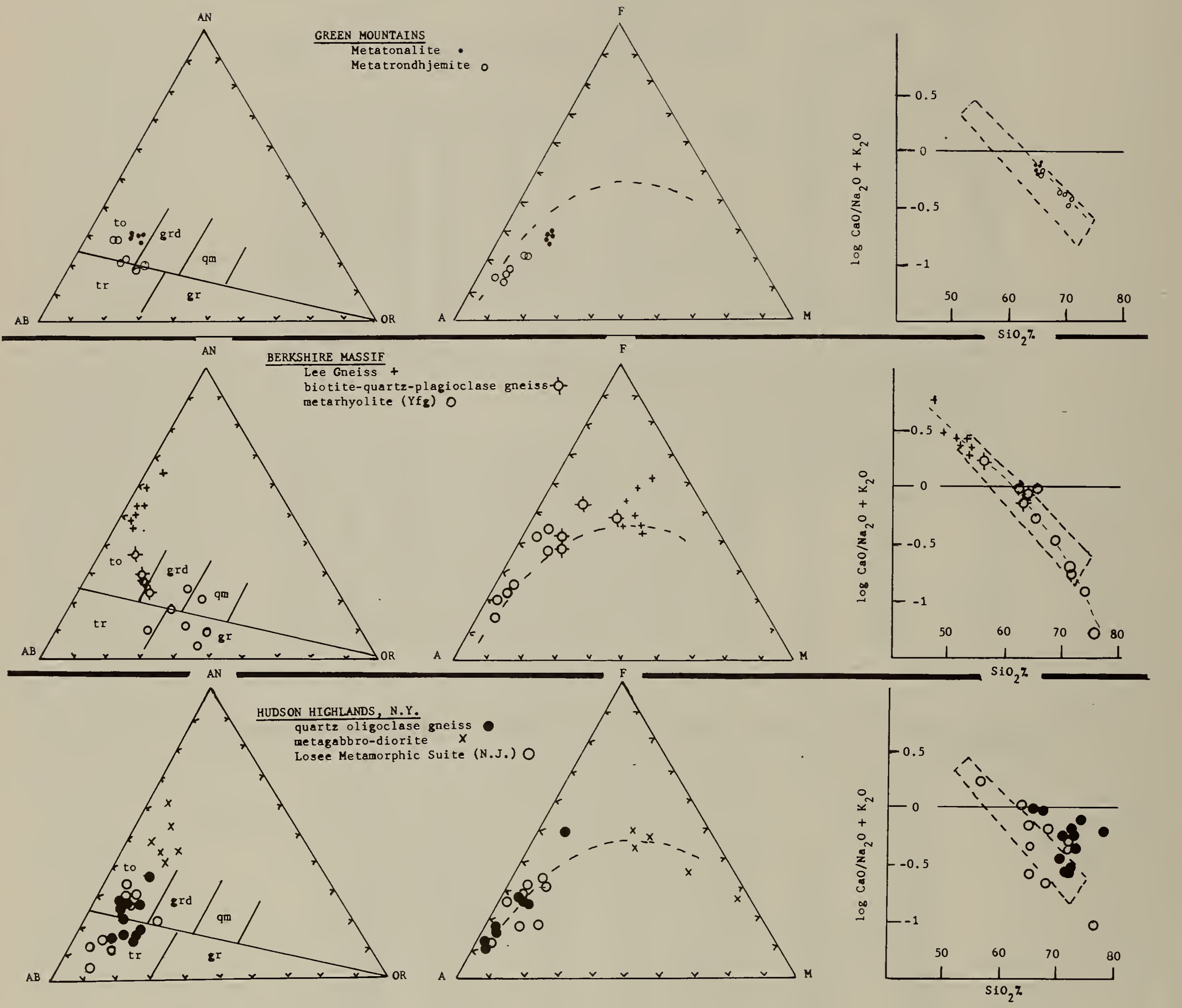



Figure 7. Selected chemical characteristics of some pretectonic metavolcanic rocks from the Grenvillian basement rocks of the northern Appalachians. Losee Metamorphic Suite from Drake (1969, 1984), all other Ratcliffe (unpub. data). AN, AB, and OR - normative anorthite, albite, orthoclase diagram showing classification scheme of O'Connor, 1965; to = tonalite, tr = trondhjemite, grd = granodiorite, qm = quartz monzonite, gr = granite; AFM - $Na_2O$ = $K_2O$, FeO total iron, Mgo, dashed line shows boundary between tholeiitic (above) and calc-alkaline suites (below) from Irvine and Baragar (1971); alkali-lime versus silica diagrams shows reference field for calc-alkaline andesite to felsic rocks after Brown (1982).

This previously unrecognized group of rocks in the Green Mountains (Yt) is also similar to K-feldspar-poor metatrondhjemite and metatonalites present in the Adirondack Highlands, especially in the southern and eastern Adirondacks as described by McLelland and others (in press), and reported by them to be approximately 1.3 Ga old based on U-Pb zircon ages. It is important to note that this Yt suite present as a discrete belt of rocks in the central Green Mountains massif, is absent from the southern Green Mountains southward to the Hudson Highlands where similar, but not necessarily correlative rocks, reappear as the quartz plagioclase gneiss and associated gabbroic rocks east of the Canopus fault system, and as the Losee Metamorphic Suite farther south in New Jersey (Drake, 1984).

In the Green Mountains, the age and stratigraphic position of the tonalite-trondhjemite gneiss belt is uncertain. In some respects the amphibolite-rich gneisses and dacitic rocks are similar to volcanics in the very thick Washington Gneiss section of central-western Massachusetts. The rusty quartzite and impure calc-silicate rocks that immediately overlie the tonalite-trondhjemite gneiss belt may be thinned equivalents of the pelitic-quartz-rich facies of the Washington Gneiss. Alternately, other explanations not withstanding, the tonalite- trondhjemite gneiss belt may unconformably underlie the biotite quartz plagioclase gneiss along a prevously unrecognized unconformity.

Late Proterozoic and Paleozoic cover sequence rocks

Cover sequence rocks overlie basement in many areas with unconformity as shown by the special contact in figure 1. The nature of the cover sequence rocks differ markedly in different lithotectonic units across the area. However, these variations are progressive rather than abrupt, except for the Whitcomb summit thrust, across which the stratigraphic succession is markedly different.

These two trips will present an opportunity to see representative sampling of cover sequence rocks from east to west exposed at or near the unconformity with basement. Quartzofeldspathic cover rocks of the Dalton Formation are well represented at the type section on and near Day Mountain (Stop 1, A-1) and at the western margin of the Green Mountain (Stop 9, A-1). Exposures of cover sequence rocks on Hoosac Mountain and in the Sadawga-Rayponda domes will be seen on trip A-1, stops 3 and 4, and on trip B-1, stops 1 and 2.

Cover sequence rocks of probable Late Proterozoic through Early Cambrian age form part of the base of the westward-thinning, transgressive clastic sequence developed on the eastern margin of North America following rifting. Within the area of these trips, the principal units are lithostratigraphic subdivisions of the Dalton Formation and its lateral, but partly older, eastern equivalent the Hoosac Formation.

The simplified nomenclature used on figure 1 shows three different sequences, following the usage on the Massachusetts State map. These are the Dalton Formation and Cheshire Quartzite (CZd-Cc) in the west, the autochthonous Hoosac section that overlies rocks of the Hoosac slice (CZhd) and the allochthonous Hoosac above the Hoosac summit thrust (CZh) to the east.

Important west to east facies changes within this belt are diagramatically shown in figure 8. Coarse conglomeratic, arkosic conglomerates (CZdc) as seen at A-1, stop 1, pass upwards into feldspathic quartzites and flagstones typical of the Dalton Formation (CZd) as a whole. Two principal horizons of dark biotite schist (CZdb) appear within the section in the Berkshires and on the southern and western margins of the Green Mountains. To the east, the percentage of dark phyllitic rocks increases as shown for the Woodford-Stratton Mountain sections, with decreasing abundance of typical (CZd) feldspathic quartzites. On Hoosac Mountain, and in the area east of the Searsburg thrust from Mount Snow south to Hoosac Mountain the basal units of the Hoosac Formation contain locally excellent Dalton-like pebbly-quartz conglomerate as well as a distinctive albite-rich granofels and a gneiss-boulder unit (CZhdc) (A-1, stop 4). Locally, black phyllite and tan weathering vitreous quartzite-similar to CZdb and CZd, respectively--are present beneath a discontinuous, beige and salmon pink weathering dolostone (CZhm) (B-1, stops 1, 2). Above or replacing this locally are discontinuous greenstones (the Turkey Mountain member of the Hoosac Formation (CZhv) of Skehan, 1961). Either above a zone of black carbonaceous phyllite or schist (CZhb) or directly resting on albite granofels (CZhab) is a coarse-grained, lustrous muscovite-paragonite(?)-chlorite-chloritoid large-garnet schist (CZhgt). Laterally, this garnet schist is replaced by or contains beds of pebbly quartz conglomerate. On Mount Snow and on the ridges south of Route 9 in the Readsboro quadrangle, these units are overlain by a thick section of fine-grained, coaly-black, lustrous, biotite-muscovite-albite-carbonaceous quartz-phyllite (CZhbc) locally containing small-garnet bearing, dark-gray, chlorite-chloritoid-muscovite phyllite. Above this unit is a second greenish-gray, fine-grained, lustrous, chlorite-chloritoid large-garnet-quartz schist (CZhg). This second garnet schist may be equivalent to the Rowe Formation or a second garnet schist within the Hoosac Formation.

## COVER SEQUENCE ROCKS

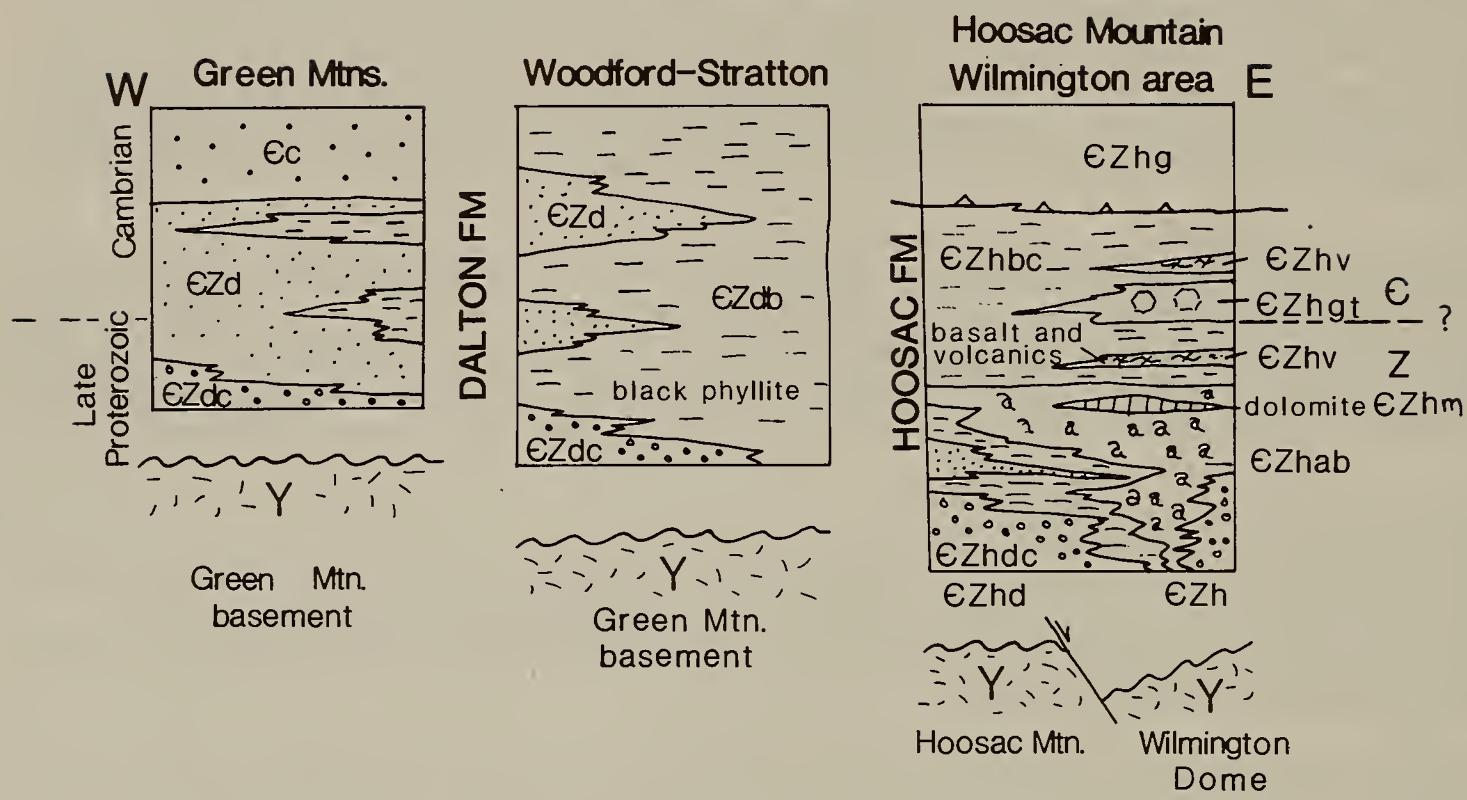



Figure 8. Generalized W to E facies variation diagram showing nomenclature for Upper Proterozoic and Lower Cambrian cover sequence rocks

The Hoosac section resting on the east side of the Rayponda dome, i.e. above the Bull Hill Gneiss (of Doll and others, 1961) contains a greater abundance of dark, coarse albite muscovite schist, several zones of hornblende-epidote greenstone and no beds of Dalton-like quartzite or black carbonaceous phyllite.

In remapping the Mount Snow and Readsboro quadrangles (part of Skehan's 1961 Wilmington 15- minute quadrangle) we have reassigned the lithologic units previously assigned to the Cavendish Formation of Doll and others, 1961, i.e. the Searsburg Conglomerate, Readsboro Formation, and Heartwellville Schist of Skehan (1961) to the Hoosac Formation based on the fact that the Hoosac Formation, as defined on Hoosac Mountain, contains all of the units present in the Cavendish Formation (Ratcliffe, 1979). The basal units of the Hoosac Formation on Mount Snow and environs (CZdc, CZhab) are roughly equivalent, therefore, to the Tyson Formation (of Doll and others, 1961), except that albitic granofels typical of the Hoosac Formation both underlies and overlies lithologies typical of the Tyson. Two aluminous-large-garnet schists, both locally containing chloritoid, appear within the section here mapped as the Hoosac Formation. The lower aluminous schist unit (CZhgt) is in the position of the Gassetts Schist member of the Cavendish Formation of Doll and others (1961), the second appears above coaly black schist and phyllite (CZhg) and may or may not be equivalent to the Rowe Formation (or Pinney Hollow Formation).

### PROTEROZOIC STRUCTURES

From Connecticut northward, Middle Proterozoic folds trending generally east-west are common in basement massifs. Locally axial traces of these folds have been mapped in Massachusetts (see A-1, stop 1). Similar kinds of folds having steep axial surfaces are present in the Green Mountains. These folds are responsible for the gross distribution of the map units in the Green Mountains as illustrated in figures 5 and 6. Fold axes of Proterozoic folds are only locally well known, but tend to plunge at gentle angles east or west within the east-west axial surfaces (see B-1, stop 9). Several generations of Proterozoic folds are known to exist, and the latest folds are well illustrated in migmatitic gneisses where thin pegmatite and aplite is intruded parallel to the axial surfaces of the latest folds. The regional distribution of these folds is presently not known. Post tectonic granitoids such as the coarse pegmatites and the 960 Ma old rapakivi granites (Ysg-Ybh) do not contain the Proterozoic structures, setting an upper limit on Grenvillian penetrative deformation. Sillimanite plus microcline-perthite and locally hypersthene have been found in members of the paragneiss suite, which coupled with the widespread occurrence of coarse brown hornblende suggests that Middle Proterozoic metamorphism of hornblende granulite grade was attained.

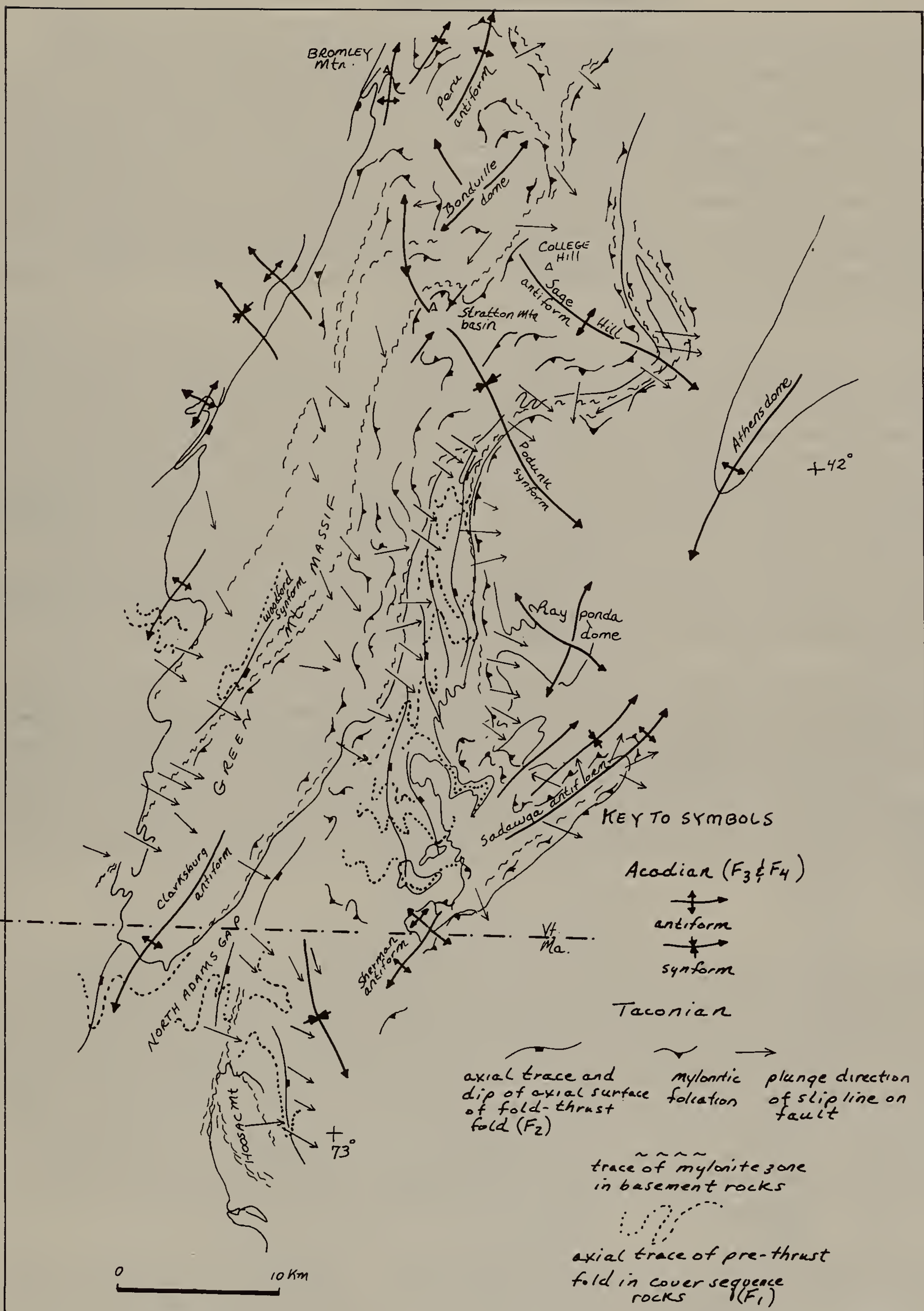



Figure 9. Simplified tectonic map showing Taconian and Acadian structures.

## PALEOZOIC STRUCTURES

The Green Mountains and Berkshire massif have been deformed during both the Taconic and Acadian orogenies. The extent and intensity of the deformations associated with both these events is a focus of our investigations. Unraveling the complexities of each is not easy, and at present we have only preliminary results. Previous studies in the Berkshire massif (Ratcliffe and Harwood, 1975; Sutter and others, 1985) have shown that the culminating effect of the Taconic orogeny involved large-scale thrust faulting of basement rocks over previously metamorphosed and folded miogeoclinal rocks. This deformation has been ascribed to collisional tectonics in which an accretionary prism overlying an oceanward-dipping subduction zone overrode the miogeoclinal wedge as described by Stanley and Ratcliffe (1983, 1985). Effects of this closure on Proterozoic basement rocks and their cover include:

(1) widespread low angle thrust faulting of imbricate slices of basement in the Berkshire massif
(2) development of mylonitic ductile deformation zones in basement rocks in which hornblende and biotite were reequilibrated syntectonically.

Samples of texturally reequilibrated biotites and hornblende that grew near the Ar closure temperatures for these two minerals, *i.e.*, lower garnet zone and staurolite zone respectively, have yielded 40Ar/39Ar plateau spectra indicating formation of these minerals and the mylonites at about 465 Ma, or Taconic ages (Sutter and others, 1985). Mylonitic ductile deformation zones form the boundaries between major lithotectonic units. Within the area of the Green Mountains covered by these trips, thrust faults and narrow semi-ductile deformation zones similar to these found in the Berkshire massif have been mapped as shown on figures 1 and 9. Lineations and mylonitic foliation within the basement rocks plunge or dip southeast but both are locally highly folded. A compliation map suggests that slip on these faults was predominantly to the northwest. Based on preliminary 40Ar/39Ar data discussed in the roadlog for trip B-1, these faults are probably Taconian, although none of the faults has been directly dated. Imbrication of basement along the Wilmington fault is also interpreted as Taconic. It is important to note that these faults along the eastern margin of the Green Mountains disrupt and crosscut earlier sets of isoclinal folds (F1 folds of figure 9) having a prominent axial planar schistosity, as do the faults bounding the Berkshire massif.

The penetrative Paleozoic fabric in the Green Mountains is highly folded. Broad areas along the central, north-south spine of the massif have subhorizontal foliation and thrust faults, forming foliation domes and basins (fig. 9). Refolds of the mylonitic foliation trend N. 25° to N. 5° E. and N. 30° to 50° W. The northeast trending folds exhibit excellent crenulation cleavage having subvertical to moderately steeply southeast-dipping axial-surfaces. This general trend forms the Green Mountain-phase folds responsible for the south-plunging antiformal closure of the Green Mountains. The northwest-trending folds rarely exhibit any cleavage, except in the schistose rocks east of the Berkshire massif and Green Mountains. Both the northeast and northwest trending folds are probably Acadian (fig. 9).

## THERMOCHRONOLOGY

In general all types of K/Ar mineral dates from metamorphic rocks should be interpreted as the time of cooling to a temperature equal to the argon closure temperature for that mineral for a particular rate of cooling. In general, the slower the rate of cooling, the lower the argon closure temperature for that mineral. In greenschist-facies rocks, the common minerals dated by the K/Ar techniques are biotite and muscovite. Although good argon diffusion data are lacking for natural biotites and muscovites, empirical studies show that for rapid cooling rates (i.e., contact metamorphism), argon is quantitatively retained in muscovite below about 350°C and in biotite below about 300°C. For slow cooling rates (i.e., regional metamorphism), the temperatures are about 320°C and 260°C, respectively. For these minerals to be separable from a prograde metamorphic mineral assemblage in a pelitic unit, that unit generally has been metamorphosed to biotite-garnet grade. The best approximation of the temperature needed to reach this grade in an aluminous metasediment is on the order of 400-450°C for moderate pressures.

Two conclusions are clear concerning the meaning of K/Ar dates on muscovite and biotite separated from prograde mineral assemblages in aluminous rocks. One is that muscovite should record an older apparent age than its coexisting biotite (this conclusion is generally true from empirical observations). The other is that both minerals must cool through a significant temperature to reach their respective closure temperatures. For slow cooling rates (regional metamorphic terrances), this cooling can last a significant amount of geologic time, and, therefore, the K/Ar dates can be significantly younger than the time of formation of a mineral, especially for biotite. For instance, if a pelitic sedimentary rock underwent prograde metamorphism to a temperature of 450°C, 465 m.y. ago and then cooled slowly at an average rate of 5°C per million years, muscovite and biotite would be expected to record apparent ages of 439 m.y. and 427 m.y., respectively, if each mineral retained all its radiogenic argon once it reached its

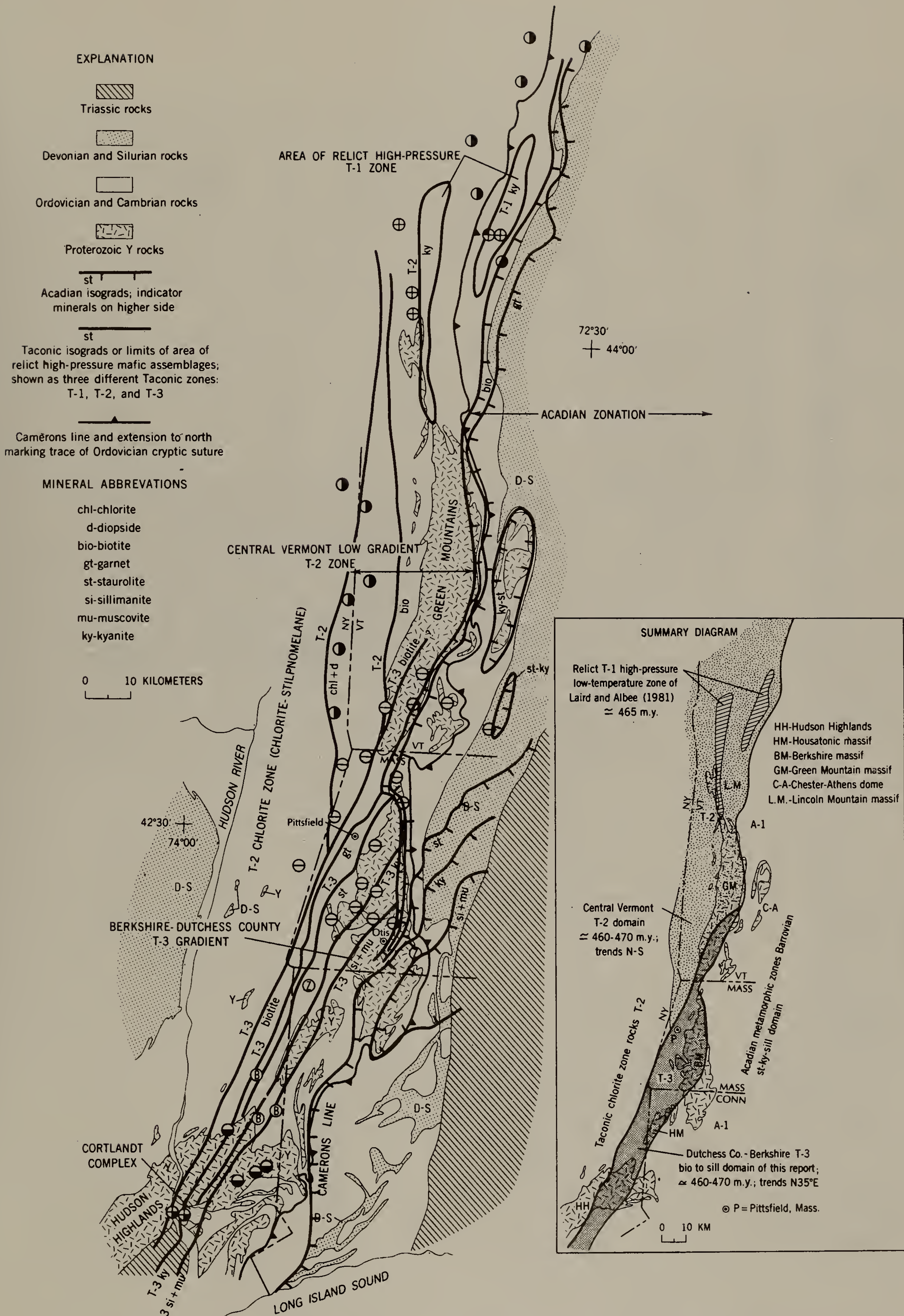
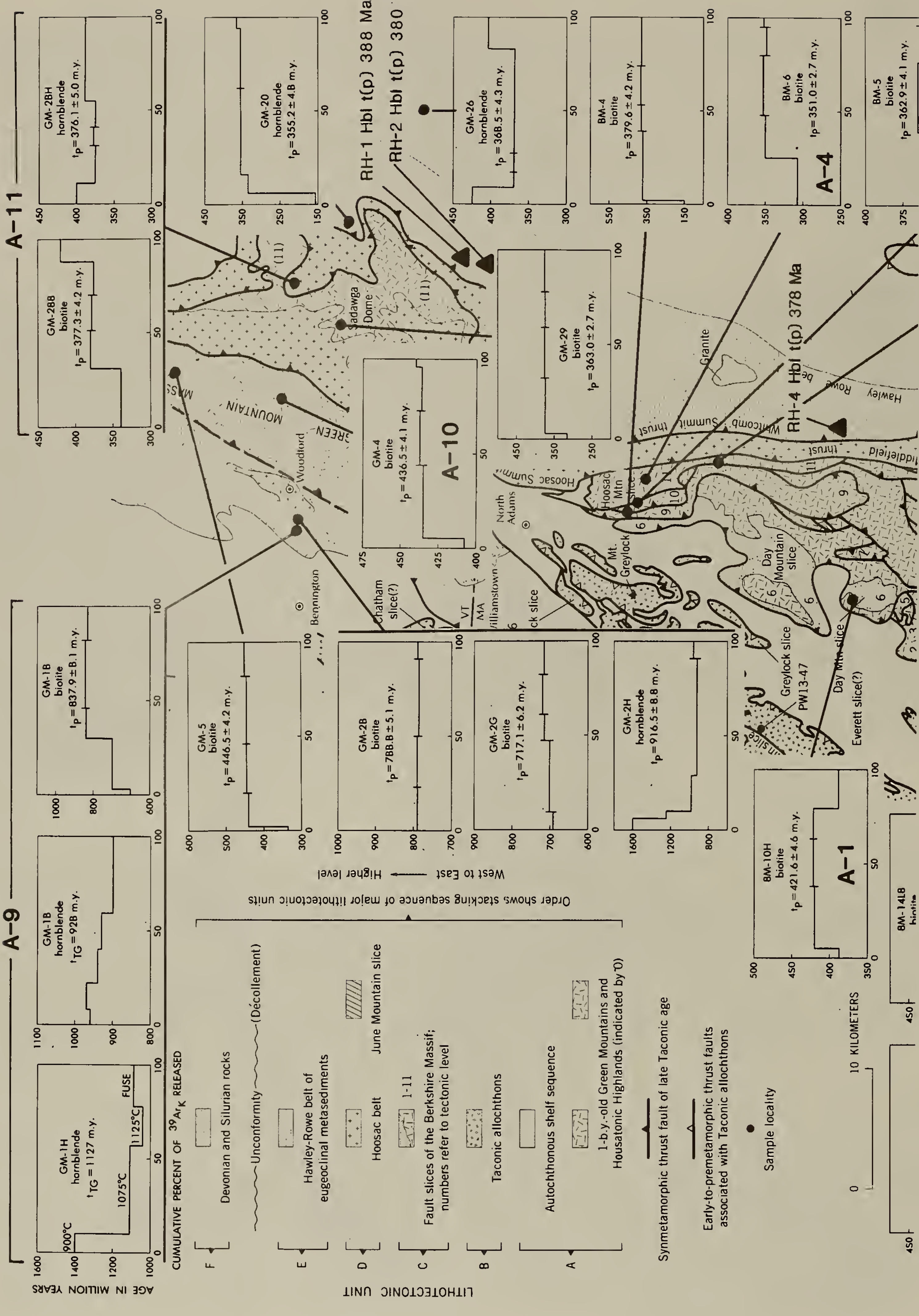



Figure 10. Suggested Taconian and Acadian metamorphic zones in western New England based on 40Ar/39Ar data reproduced from Sutter and others, 1985.



A-9
A-11
A-10
A-4
A-1
GM-1H hornblende $t_{TG}$ = 1127 m.y.
900°C
1075°C
1125°C
FUSE
GM-1B hornblende $t_{TG}$ = 928 m.y.
GM-1B biotite $t_p$ = 837.9 ± 8.1 m.y.
GM-28B biotite $t_p$ = 377.3 ± 4.2 m.y.
GM-2BH hornblende $t_p$ = 376.1 ± 5.0 m.y.
GM-20 hornblende $t_p$ = 355.2 ± 4.8 m.y.
GM-5 biotite $t_p$ = 446.5 ± 4.2 m.y.
GM-2B biotite $t_p$ = 788.8 ± 5.1 m.y.
GM-2G biotite $t_p$ = 717.1 ± 6.2 m.y.
GM-2H hornblende $t_p$ = 916.5 ± 8.8 m.y.
GM-4 biotite $t_p$ = 436.5 ± 4.1 m.y.
GM-29 biotite $t_p$ = 363.0 ± 2.7 m.y.
GM-26 hornblende $t_p$ = 368.5 ± 4.3 m.y.
BM-4 biotite $t_p$ = 379.6 ± 4.2 m.y.
BM-6 biotite $t_p$ = 351.0 ± 2.7 m.y.
BM-5 biotite $t_p$ = 362.9 ± 4.1 m.y.
BM-10H biotite $t_p$ = 421.6 ± 4.6 m.y.
BM-14LB biotite
RH-1 Hbl t(p) 388 Ma
RH-2 Hbl t(p) 380 Ma
RH-4 Hbl t(p) 378 Ma
AGE IN MILLION YEARS
CUMULATIVE PERCENT OF $^{39}Ar_K$ RELEASED
LITHOTECTONIC UNIT
Devonian and Silurian rocks
Unconformity (Décollement)
Hawley-Rowe belt of eugeoclinal metasediments
Hoosac belt
June Mountain slice
Fault slices of the Berkshire Massif; numbers refer to tectonic level
Taconic allochthons
Autochthonous shelf sequence
1-b.y.-old Green Mountains and Housatonic Highlands (indicated by 0)
Order shows stacking sequence of major lithotectonic units
West to East —— Higher level
Synmetamorphic thrust fault of late Taconic age
Early-to-premetamorphic thrust faults associated with Taconic allochthons
Sample locality
10 KILOMETERS
GREEN MOUNTAIN MASS
Woodford
Bennington
Chatham slice(?)
VT
MA
Williamstown
North Adams
Mt. Greylock
Hoosac Mtn
Hoosac Summit
Whitcomb Summit thrust
Middlefield thrust
Hawley Rowe belt
Granite
Sadawga Dome
Day Mountain slice
Greylock slice
PW13-47
Day Mtn. slice
Everett slice(?)

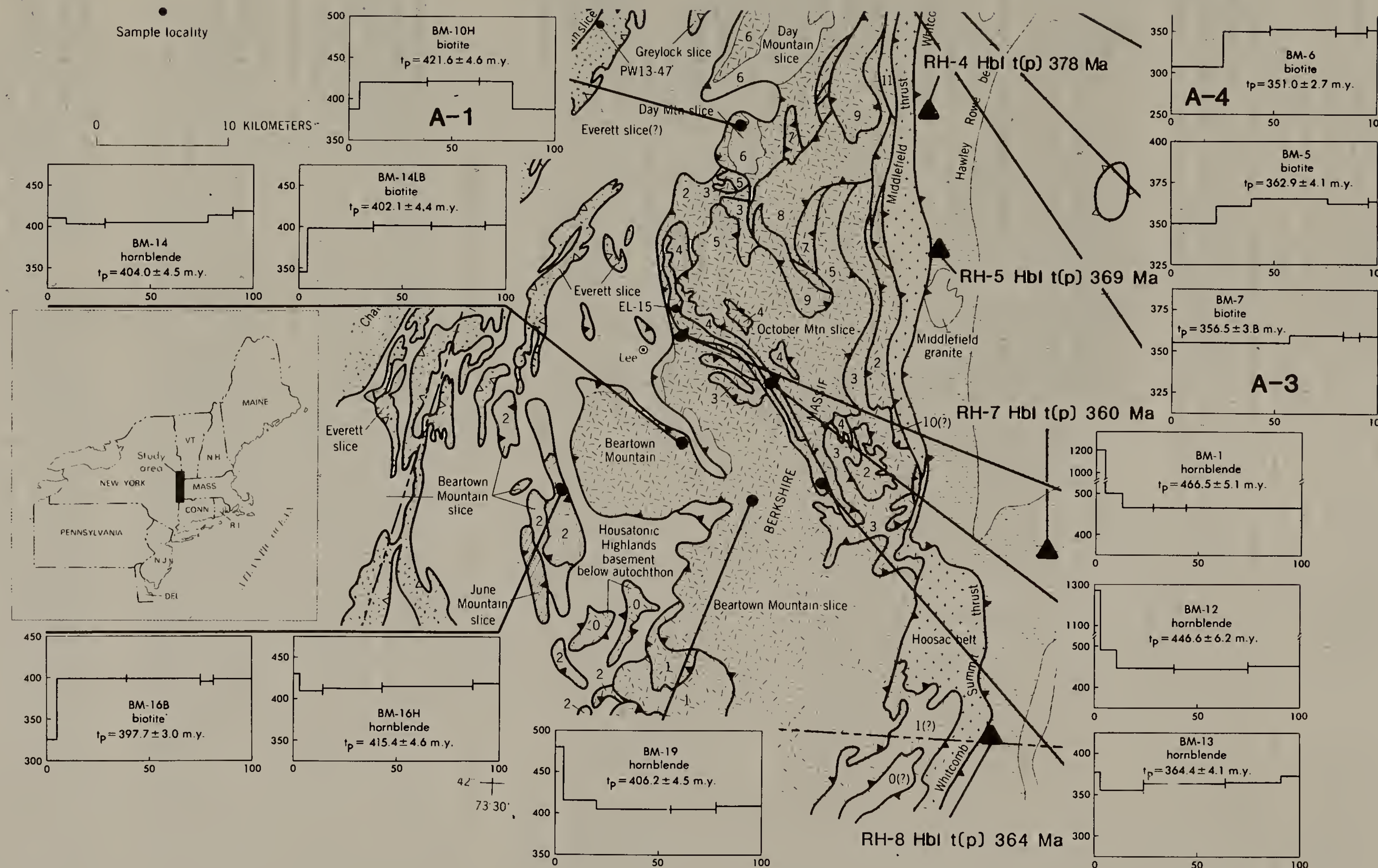



Figure 11. 40Ar/39Ar release spectra for samples from the southern Green Mountains and northern Berkshire massif from Sutter and others, 1985, additional data from Sutter and Hatchk, 19--, in the Rowe-Hawley zone shown by solid triangles.

closure temperature. If the time-averaged cooling rate was as low as 3°C per million years (typical for Taconian domain T-3) and if the temperature maximum was 500°C, 465 m.y. ago, coexisting muscovite and biotite would be expected to record argon closure ages of about 405 m.y. and 385 m.y., respectively. These apparent ages could be construed to be "Acadian" on face value but really represent slow cooling from a "Taconian" metamorphic maximum. In general, the apparent K/Ar dates of biotite and muscovite should decrease as metamorphic grade increases in a prograde sequence. According to this reasoning, the closest approximation to the time of formation of each mineral will be the apparent date measured for that mineral at the lowest metamorphic grade, but even that date can be (and commonly is) significantly younger than the mineral's time of formation. Thus, mica K/Ar dates from rocks above biotite-garnet grade should not generally be used to estimate times of mineral growth unless the cooling rate following formation is known to have been rapid.

The use of the 40Ar/39Ar age spectrum technique in thermochronology is well established. In short, the argon present in the mineral (both radiogenic and neutron-produced) is extracted in the laboratory in a series of progressively higher temperature steps and apparent ages are calculated for each and graphically displayed as an age spectrum diagram. If the apparent ages of all temperature steps are not analytically the same, the spectrum is said to be discordant. Age plateaus are defined when multiple temperature steps, together representing more than 50 percent of the total argon in the sample, yield the same age even though the age spectrum as a whole is discordant. The age spectra of biotite and hornblende from regionally metamorphosed rocks are generally somewhat discordant but usually form age plateaus. The discordance of the age spectrum is often caused by small amounts of argon loss or argon gain (excess argon). The plateau ages are taken to be the best approximation of the time of cooling to the argon closure temperature for that mineral. For the slow cooling rates often encountered in regional metamorphic terranes, the closure temperature for hornblende is about 480-500°C, and that for biotite is about 260-280°C. When coexisting biotite and hornblende from a prograde metamorphic mineral assemblage are dated by the 40Ar/39Ar age spectrum technique, the plateau age of the hornblende should be older than the plateau age of the biotite unless cooling between 480°C and 260°C was very rapid. Again, hornblende and biotite plateau ages should decrease as metamorphic grade increases, and the difference in plateau ages of coexisting biotite and hornblende should reflect the rate of cooling between their respective argon closure temperatures. In addition, for hornblende to be a separable mineral phase from a prograde metamorphic rock, that rock generally has been metamorphosed to at least garnet-staurolite grade. The best estimate for the minimum temperature necessary for this metamorphic grade is about 500°C. Therefore, in garnet-staurolite-grade rocks that contain prograde hornblende, that hornblende grew at a temperature very near to its argon closure temperature. Thus, to make the best approximation to the timing of arrival at the thermal maximum in a prograde metamorphic sequence, 40Ar/39Ar age spectra should be measured on hornblendes from rocks in the garnet-staurolite zone. In addition, hornblende plateau ages from higher grade rocks, together with petrologic data, can be used to estimate the rate of post-metamorphic cooling to the argon closure temperature of hornblende.

## METAMORPHISM AND REGIONAL DYNAMOTHERMAL DOMAINS

Three Taconian metamorphic/structure zones (T-1, T-2, T-3 in fig. 10) are recognized in the polydeformed belt of western New England. On these trips we are dealing largely with the structural and metamorphic overprint fabrics associated with T-3 zone and the boundary with Acadian metamorphic and structural overprinting to the east. The T-3 zone is a discrete structural and petrographic zone of late-Taconian thrust-faulting and prograde Barrovian metamorphism characterized from New York to the northern end of the Berkshire massif by a steep metamorphic gradient from biotite to sillimanite + muscovite grade. This high-gradient zone, contains throughout, metamorphic textural features indicative of Taconian polymetamorphism and multiple-phase Taconian structure, i.e., the most complex part of the Taconian metamorphic belt. Rock textures and mineral assemblages in this belt last equilibrated in the T-3 event in which mobilization of basement and large-scale sialic thrust slivers were imbricated upwards into the already accreted allochthonous terrains of the Taconics. Ar/Ar hornblende ages from the southern Berkshire massif and from the Hudson Highlands have been shown to approximate the development of T-3 aged mylonitic rocks in basement rocks remobilized during this late Taconian event.

In the area of these field trips the T-3 metamorphic gradient from garnet zone to staurolite-kyanite zone disappears beneath a nearly north-south-trending Acadian overprint of biotite to upper garnet zone. The distribution of pre-Acadian T-3 zonation is unknown. The presence of zoned unconformity garnets within the cover sequence rocks near Jamaica, Vt. (Karabinos, 1984, and this volume) and in the Chester and Athens domes (Rosenfeld, 1968, and this volume) indicates that the Taconian garnet zone (probably T-3) extends northward under the Acadian overprint along strike of the isograd projected from the south. However, no relict Taconian mineral assemblages of staurolite-kyanite or higher grade are recognized in the cover rocks of the Sadawga-Rayponda or Chester-Athens

domes, thus leaving open the question of the regional extent of these higher T-3 zones. The high T-3 zones are thought to reflect higher tectonic loads resulting from greater sialic imbrication from Vermont southward to New York. If this interpretation is correct, then T-3 zonation greater than garnet zone may have closed to the north and never been present in the cover rocks of Vermont. A corollary of this is that net amount of basement imbrication descreases northward, and that an increasing amount of net-throw was transferred northward onto the lower, Champlain thrust.

From New York to the central part of the Berkshire massif T-3 age thrust faults have sliplines indicative of movement from northeast to southwest, *i.e.*, faults have a component of right-lateral strike-slip motion. From the northern Berkshire massif northward, similar basement overthrusts and the Champlain thrust have movement from the southeast to the northwest or a component of left-lateral strike-slip motion. Ratcliffe (1988) has suggested that this variation as well as the distribution of the Taconic allochthons can all be explained by the accretionary- collapse of an eastward-facing Berkshire promontory during the Taconian collisional event (Stanley and Ratcliffe, 1985). The remains of the Berkshire promontory are now present in the accumulated imbricate slices of sialic basement rocks making up the Berkshire massif and recumbently folded areas of the Manhattan Prong. The promontory, therefore has been inverted to produce a large structural salient verging to the west.

The change in thrust fault movement to the northwest and the lesser extent of internal thrust faulting within the Green Mountains, as opposed to that in the Berkshire massif, is consistent with that model. On these field trips we will try to evaluate many of these concepts by discussion of both weaknesses and strengths of these arguments.

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## ROAD LOG TRIP A-1

Assembly point 600 feet south of intersection of routes 8 South and 9 in the town of Coltsville, 2 miles east of Pittsfield, Mass. Assemble in parking lot of supermarket just south of light on the east side of road that is the extension of route 8 to the south.

Assembly time 8:30 am, Friday October 14th
If you should arrive late, follow roadlog to first stop and follow flagging down and up slope to first locality. We will be at Stop 1 until 9:15 am.

Mileage (cumulative)

0.0 Intersection route 9 and 8 south Coltsville turn right (east) on 8 south and follow 1.3 miles to Dalton.

1.3 Turn right onto South Street at Crane Paper Company Museum Turn left on Grange Hall Rd. drive up hill 0.9 mile to point just past mailbox and driveway to the right

3.0 STOP 1. Day Mountain unconformity and Taconian staurolite grade rocks.

Unconformity of Dalton Formation conglomerate on Middle Proterozoic gneiss. Day Mountain Pittsfield East quadrangle. Starting from road walk north into brook to exposures of hornblende and biotite layered gneiss showing subvertical Proterozoic layering. Biotite from this rock (BM10H) gives a Ar/Ar plateau age of 421.6 Ma (fig. 11). Because aluminous rocks underneath the Day Mountain slice at the foot of North mountain in Dalton contain staurolite, temperatures exceeded the closure temperature of biotite here and the 422 Ma age is interpreted as a Taconian cooling age. Follow gneiss outcrops diagonally (NNE) up the slopes to large ledges of gneiss and overlying conglomerate. This is one of the two best exposures of the unconformity between the Dalton Formation and the basement gneiss in western Massachusetts. B. K. Emerson (1899) described this locality as his Dalton Club House locality. Few words are necessary at this exposure. Walk along the low cliffs and locate the unconformity and the angular discordance. This outcrop is figured in Ratcliffe and Zartman (1976); the geology in Ratcliffe (1984). Return downhill to left turn onto South Street Turn left (0.9 mi) at Division Street
Turn left on Washington Mtn. Rd., head up hill and bear right at Y park near bend in road to right

7.9 STOP 2. Thrust fabric in gneiss and discussion of fault structures in Pittsfield East quadrangle.

Outcrops by road are mylonitic, augen gneiss derived from the Tyringham Gneiss, a Middle Proterozoic augen gneiss interpreted as a syntectonic Middle Proterozoic biotite granite. Structures and textures in the Tyringham regionally show that it participated in the intense Grenvillian deformation that preceeded the deposition of the Dalton Formation (Ratcliffe and Zartman, 1975). In this regard the Tyringham differs from the 950 to 900 Ma old Stamford Granite Gneiss and granitoids to be seen later, because the latter lacks a Middle Proterozoic fabric. Granitic gneiss at this outcrop are thus twice deformed or 2x augen gneiss, as opposed to 1 cycle augen gneisses derived from the Stamford seen at Stop 2. Thrust fault fabric strikes northwest and dips southwest parallel to the contact with the underlying Dalton Formation. In the northern Berkshire massif numerous thrust faults are marked by similar zones of mylonite. These faults stack up to the north-northeast. Paleozoic shortening through structural overlap amounts to about 47 km in the 10 km long section illustrated in Ratcliffe (1984). The complex, nested-thrust fault style seen in the Pittsfield East quadrangle is typical of the Berkshire massif as a whole and atypical of the structures developed in the Green Mountains massif. In the Green Mountains few if any through-going thrust faults are recognized. This difference in tectonic style is an important contrast to be developed on this field trip and Saturday's trip through the Green Mountains. The complex slivering of basement and cover rocks characteristic of the Berkshire massif continues northward into the Wilmington area and further northward to the eastern margin of the Green Mountain massif at Jamaica, Vt. From this point north regionally important thrust faults may re-enter the Green Mountain basement as suggested by Karabinos (1987), although Ratcliffe suspects that the bulk of the faulting is in and around the repeated slices of basement rocks east of Jamaica and in the Chester-Athens domes, along the Cobbe Brook fault of Karabinos (1984).

Turn around and head downhill to turn right on Division Street and continue 1.6 miles to traffic light, continue across and follow road north to Rt. 8 south. 1

2.3 Turn right on Rt. 8 south, proceed E through town of Dalton then turn left on routes

14.2 8A and 9. Follow 8A and 9 to Windsor 6.3 miles and turn left on route 8A, follow route 8A (4.4 mi) to the intersection with Rt. 116 and park

24.9 STOP 3. The Cuddleybunny outcrop. Mylonitic fabrics and Acadian overprint.

This outcrop in the Windsor quadrangle was previously described by Norton (1975), who mapped this quadrangle (Norton, 1967). These road cuts of biotite plagioclase gneiss are all strongly tectonized and contain strong Paleozoic overprint and mylonitic foliation dipping northeast beneath the overlying Stamford Granite Gneiss exposed at the east end of the cut. Norton mapped these rocks as forming the northeastern upright limb of a nearly recumbent antiform cored by biotite plagioclase gneiss and unconformably overlain by a transitional facies of schistose Hoosac, a biotite muscovite garnet schist and thin discontinuous vitreous quartzite. Within the roadcut one belt of muscovitic biotite-garnet schist is interpreted by Norton as a folded fault sliver of Hoosac Formation, that in turn is overridden to the east by mylonitic Stamford Granite Gneiss.

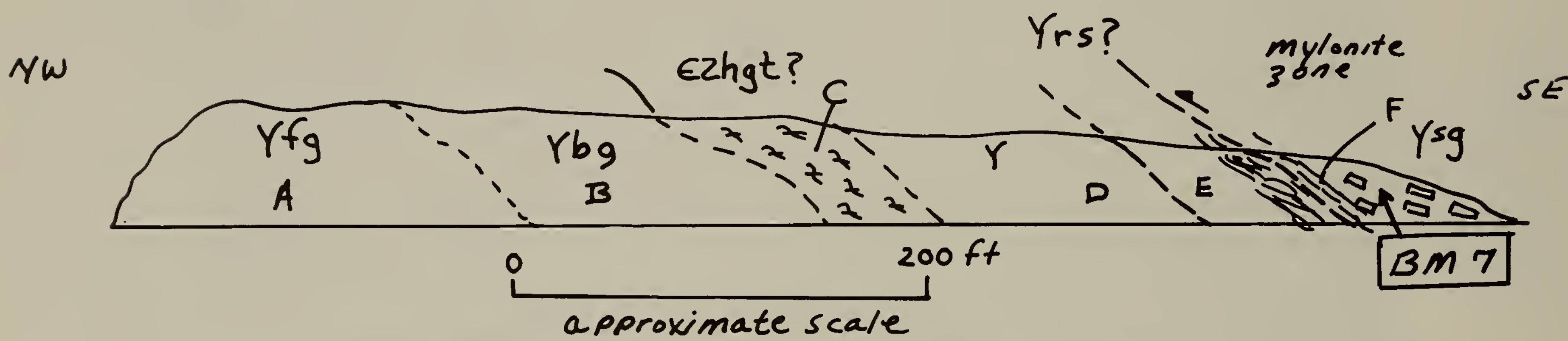


FIgure 12. Sketch of roadcut at Stop A-3 showing location of analytical samples in table 6.

Chemical analyses of rocks from the cut (fig. 12) are given in table 7. Light colored felsic gneiss (Yf, samples A and D) well-layered biotite-quartz-plagioclase gneiss (Ybg, sample B) dark biotite mylonitic gneiss (Ybm, sample E) are typical of the metavolcanic plagioclase quartz gneiss present in the Berkshire massif, comparable to the calc-alkaline suite shown in table 6. Garnet schist possibly of the Hoosac Formation is sample C. Two samples F1 and F2 are fine-grained dark mylonitic and coarser Stamford Granite Gneiss. Dispite the markedly differ appearance of the dark and light colored layers their major element, and REE abundances are identical suggesting mylonitization was isochemical involved no differential volume change (between the two layers).

Biotite sample BM7 from mylonitic Stamford Granite Gneiss (tp = 356.5 Ma) is Acadian and indicates that Acadian metamorphism at least of biotite grade affected these rock. The progression in age from samples from BM4, BM6, BM5 (379 to 356 Ma) suggests that all of these samples represent cooling ages from Acadian remetamorphism in the biotite through garnet zone. Garnets in the muscovite-biotite-plagioclase-quartz schist (C) have inclusion rich cores, and clear euhedral overgrowth rims. The cores probably are Taconian, the overgrowvths Acadian. Continue E on route 8A and 116 0.5 mi and turn left on Center Rd., in 2.9 mi take left at intersection following sign to Savoy Mtn. State Forest and 0.2 further bear right

29.1 At Y intersection; following sign to Savoy Mtn. State Forest; take left "Y" on Burnett Rd. and then right onto Florida road past entrance to Savoy Mt. Forest, go past North Pond and look for dirt road turning left from sharp right hand bend in road pull into dirt road to clearing

31.1 STOP 4. Stamford Granite Gneiss and unconformable cover sequence, north end of Hoosac Mountain, North Adams quadrangle.

Table 6. Chemical analyses of biotite quartz plagioclase leucogneiss (metatrondhjemite), Jamaica and Peru quadrangles, VT, and metatonalite at Cole Pond

| Sample | 2755 | 2750 | 2754 | 3025 | 1230 | 3153 | 3016A | 3016B | 3016C | 3016D | 3016E |
|---|---|---|---|---|---|---|---|---|---|---|---|
| | metatrondhjemite | | | | | metatonalite | | | | | |
| constituent | | | | | | | | | | | |
| $SiO_2$ | 66.6 | 66.6 | 70.0 | 69.9 | 71.5 | 70.8 | 65.2 | 65.5 | 65.3 | 66.0 | 66.1 |
| $Al_2O_3$ | 18.1 | 18.1 | 16.3 | 16.4 | 15.50 | 16.0 | 17.1 | 17.2 | 17.1 | 16.6 | 16.8 |
| FeO | 1.2 | 1.1 | 0.92 | 0.52 | 0.40 | 0.56 | 2.0 | 1.90 | 1.6 | 1.8 | 1.7 |
| $Fe_2O_3$ | .9 | 1.0 | 0.57 | 1.0 | 0.94 | 0.63 | 1.23 | 1.15 | 1.6 | 1.4 | 1.3 |
| MgO | 1.05 | 1.13 | 0.77 | 0.81 | 0.46 | 0.80 | 1.82 | 1.69 | 1.90 | 1.56 | 1.45 |
| CaO | 4.18 | 4.17 | 2.99 | 2.91 | 2.60 | 2.45 | 4.52 | 4.28 | 4.60 | 4.81 | 4.97 |
| $Na_2O$ | 5.12 | 5.32 | 5.03 | 4.94 | 4.53 | 4.92 | 4.59 | 4.52 | 4.53 | 4.25 | 4.42 |
| $K_2O$ | 1.21 | 1.12 | 1.76 | 1.68 | 2.51 | 2.30 | 1.71 | 2.08 | 1.75 | 1.94 | 1.61 |
| $TiO_2$ | 0.24 | 0.26 | 0.18 | 0.20 | 0.14 | 0.18 | 0.45 | 0.39 | 0.45 | 0.23 | 0.23 |
| $P_2O_5$ | 0.12 | 0.13 | 0.08 | 0.08 | 0.07 | 0.09 | 0.15 | 0.14 | 0.15 | 0.13 | 0.12 |
| MnO | <0.02 | 0.02 | 0.02 | <0.02 | 0.02 | <0.02 | 0.04 | 0.05 | 0.05 | 0.06 | 0.06 |
| $H_2O^+$ | .48 | 0.15 | 0.20 | 0.20 | 0.44 | 0.34 | 1.3 | 0.47 | 0.53 | .52 | 0.39 |
| $H_2O^-$ | 0.21 | 0.21 | 0.17 | 0.16 | 0.08 | 0.26 | 0.08 | 0.10 | 0.08 | .04 | 0.05 |
| $CO_2$ | <0.01 | <0.01 | 0.02 | 0.01 | 0.01 | <0.01 | 0.11 | 0.13 | 0.09 | .22 | 0.25 |
| Total | 99.11 | 99.32 | 99.01 | 98.83 | 99.47 | 99.38 | 100.3 | 99.6 | 99.73 | 99.56 | 99.45 |
| CIPW Norms | | | | | | | | | | | |
| constituent | | | | | | | | | | | |
| Q | 21.8 | 21.0 | 26.7 | 28.0 | 29.9 | 27.5 | 19.6 | 19.6 | 20.0 | 21.9 | 22.2 |
| C | 1.1 | 0.9 | 0.9 | 1.4 | 0.8 | 1.2 | 0.1 | 0.4 | 0.0 | 0.0 | 0.0 |
| or | 7.2 | 6.7 | 10.6 | 10.1 | 15.0 | 13.8 | 10.2 | 12.4 | 10.5 | 11.6 | 9.6 |
| ab | 43.9 | 45.5 | 43.2 | 42.5 | 38.9 | 42.2 | 39.4 | 38.7 | 38.8 | 36.5 | 37.9 |
| an | 20.2 | 20.0 | 14.4 | 14.1 | 12.6 | 11.7 | 21.0 | 19.7 | 21.4 | 20.8 | 21.5 |
| di | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.1 | 0.9 | 0.9 |
| hy | 3.8 | 3.7 | 2.9 | 2.1 | 1.2 | 2.3 | 6.6 | 6.3 | 5.7 | 5.4 | 5.0 |
| mt | 1.3 | 1.5 | 0.8 | 1.2 | 1.0 | 0.9 | 1.8 | 1.7 | 2.3 | 2.1 | 1.9 |
| hm | 0.0 | 0.0 | 0.0 | 0.2 | 0.3 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 | 0.0 |
| il | 0.5 | 0.5 | 0.3 | 0.4 | 0.3 | 0.3 | 0.9 | 0.7 | 0.9 | 0.4 | 0.4 |
| ap | 0.3 | 0.3 | 0.2 | 0.2 | 0.2 | 0.2 | 0.4 | 0.3 | 0.4 | 0.3 | 0.3 |
| cc | 0.0 | 0.0 | 0.1 | 0.0 | 0.0 | 0.0 | 0.3 | 0.3 | 0.20 | 0.5 | 0.6 |
| Total | 100.1 | 100.1 | 100.1 | 100.2 | 100.2 | 100.1 | 100.3 | 100.1 | 100.3 | 100.3 | 100.3 |

Table 7. Chemical analyses of gneissic rocks at road cut Rt. 116 and 8A in Windsor Quadrangle

| Sample | 1508A | 1508B | 1508C | 1508D | 1508E | $1508F_1$ | $1508F_2$ |
|---|---|---|---|---|---|---|---|
| constituent | | schist | | | | Stamford | |
| $SiO_2$ | 74.50 | 65.60 | 46.70 | 76.0 | 62.8 | 66.5 | 67.2 |
| $Al_2O_3$ | 13.80 | 15.50 | 25.60 | 12.40 | 15.40 | 14.1 | 14.1 |
| FeO | 0.44 | 3.90 | 8.80 | 0.84 | 4.70 | 4.4 | 4.0 |
| $Fe_2O_3$ | 0.25 | 1.09 | 2.11 | 0.54 | 1.75 | 1.29 | 1.57 |
| MgO | 0.20 | 1.18 | 2.31 | 0.57 | 0.46 | 1.43 | 0.63 |
| CaO | 1.08 | 3.08 | 1.42 | 0.38 | 3.61 | 1.57 | 2.01 |
| $Na_2O$ | 3.49 | 4.28 | 1.33 | 3.60 | 3.38 | 2.65 | 3.12 |
| $K_2O$ | 4.93 | 2.44 | 5.51 | 4.49 | 4.43 | 4.84 | 4.67 |
| $TiO_2$ | 0.05 | 0.72 | 1.43 | 0.72 | 0.79 | 0.64 | 0.64 |
| $P_2O_5$ | 0.05 | 0.22 | 0.25 | 0.05 | 0.34 | 0.21 | 0.20 |
| MnO | 0.02 | 0.02 | 0.25 | 0.02 | 0.09 | 0.05 | 0.08 |
| $H_2O^+$ | 0.36 | 0.70 | 2.60 | 0.29 | 0.68 | 0.72 | 0.79 |
| $H_2O^-$ | 0.12 | 0.11 | 0.16 | 0.12 | 0.08 | 0.18 | 0.21 |
| $CO_2$ | 0.06 | 0.05 | 0.01 | 0.02 | 0.66 | <0.01 | 0.01 |
| Total | 99.89 | 98.89 | 98.48 | 99.44 | 99.17 | 98.58 | 99.22 |
| CIPW Norms | | | | | | | |
| consitituent | | | | | | | |
| Q | 33.2 | 21.9 | 6.7 | 36.4 | 18.2 | 25.6 | 24.7 |
| C | 1.0 | 0.9 | 16.2 | 1.1 | 0.8 | 2.2 | 0.8 |
| or | 29.5 | 14.7 | 34.1 | 26.8 | 26.8 | 29.3 | 28.1 |
| ab | 29.9 | 37.0 | 11.8 | 30.8 | 29.3 | 23.0 | 26.9 |
| an | 4.7 | 13.8 | 5.6 | 1.4 | 11.8 | 6.5 | 8.8 |
| hy | 1.1 | 8.2 | 19.1 | 2.4 | 7.4 | 9.9 | 6.8 |
| mt | 0.4 | 1.6 | 3.2 | 0.8 | 2.6 | 1.9 | 2.3 |
| il | 0.1 | 1.4 | 2.8 | 0.2 | 1.5 | 1.3 | 1.2 |
| ap | 0.1 | 0.5 | 0.6 | 0.1 | 0.8 | 0.5 | 0.5 |
| cc | 0.1 | 0.1 | 0.0 | 0.1 | 1.5 | 0.0 | 0.0 |
| Total | 100.1 | 100.1 | 100.1 | 100.1 | 100.7 | 100.2 | 100.1 |

Gray, albitic granofels and quartz pebble conglomerate of the Hoosac Formation unconformably overlies Stamford Granite Gneiss. This basal unit is well developed here as well as on the west side of the Wilmington antiform where it forms a distinctive unit previously mapped locally by Skehan (1961) as Searsburg Conglomerate. This albitic conglomerate contains clasts of pegmatitic Stamford up to 3 feet in diameter locally. Elsewhere lensoidal albitic "clasts" may be disarticulated beds producing a pseudoconglomerate. Dalton-like feldspathic quartzite and dark biotite schist (CZdbs) overlies this albitic facies, thus suggesting that the basal Hoosac is an eastern and older part of the Dalton-Cheshire transgressive quartzofeldspathic sequence. Walk along logging road to contact of basal Hoosac and Stamford at the stream. Samples of typical Stamford are located near the unconformity.

Continue north on Florida Rd., then take left on So. County Rd. in (0.9 mi), right at first Y intersection, go past Hoosac tunnel ventillation shaft, bear left at next Y intersection.

34.1 Turn right on Rt. 2, 1 mile east turn left opposite Florida Fire Dept. on Tilda Hill Rd., follow paved Rd. 2.9 miles, turn left on Turner Hill Rd., follow 1.6 miles to first right turn, follow this dirt road 1.8 mile downhill past P.O. Box 158, turn left, in 0.5 mile turn right onto narrow dirt road parallel to Beaver Brook, continue 0.6 mile to power line and park.

50.9 STOP 5. Basal albitic conglomerate of the Hoosac Formation Readsboro quadrangle.

Excellent near crops of very coarse-grained albitic granofels and conglomerate typical of the basal Hoosac can be seen at the base of the slopes leading up the powerline. Scattered crops of biotite plagioclase gneiss and probable calc-silicate gneiss are found to the north below the conglomerate. Albitic granofels passes upward into dark black to gray, garnet-biotite- plagioclase-quartz schist and rusty muscovitic garnet schist. At the crest of the hill light green, very aluminous garnet-chlorite-chloritoid-muscovite (paragonite?) quartz schist and large garnet schist assigned to the allochthonous Hoosac overlies the albitic Hoosac. This contact is interpreted as the Hoosac Summit Thrust on figure 1.

Right turn on Rt. 2. Follow Rt. 2, over crest of hill and down to hairpin turn. Excellent views of Mount Greylock to the southwest and Clarksburg Mount and the southern end of the Green Mountains to the west. The aluminous Hoosac Formation schists exposed in the road cuts form part of the allochthonous Hoosac section that lies above the Hoosac summit thrust (see map).

55.3 Turn right on Rt. 8, in 3.0 miles turn left on Middle Rd., in 1.0 mi turn right on Horrigan Rd., and in 1.3 miles take left on Lesure Rd.

Park 300 feet up road.

60.6 STOP 6. Stamford Granite Gneiss and unconformable Dalton Formation cover.

This tired, old, but excellent exposure is always willing to entertain geologists--apologies to those of you who have been here 16 times already! A U-Pb zircon age of 958+4 Ma has been obtained from samples at this spot. The nearly concordant age (Karabinos and Aleinikoff, 1988) is exceptionally important. Our mapping of the Stamford here and on Hoosac Mountain indicates that the Stamford crosscuts Grenvillian sillimanite grade (hornblende granulite facies rocks) but is not deformed in the Proterozoic, i.e. lacks the gneissic fabric characteristic of other Middle Proterozoic granitic rocks such as the Tyringham and granitic gneisses of the Green Mountains. Chemical characteristics of the Stamford are presented in table 3 and figures 5 and 13. Mafic to intermediate dikes and irregular masses within the granite suggest that gabbroic to monzonitic liquids are associated with the coarse rapakivi granite, which is a K-feldspar cumulate. A strong positive Eu anomaly characterizes the Stamford here and on Hoosac Mountain. Where mafic dikes crosscut or are in contact with the coarser grained granite, K-feldspar megacrysts, quartz and plagioclase have resorption features and a new generation of K-feldspar having rapakivi structure form in the matrix. The mafic rocks are interpreted as comagmatitic liquids that intruded and reacted with the feldspar cumulates. 150 yards up the road one of these mafic dikes crosscuts the Stamford. On the Massachusetts State Map (Zen and others, 1983) this dike is assigned to the Late Proterozoic. Chemical analyses of this dike (table 3) and comparison with Late Proterozoic basaltic volcanics in the Hoosac Formation indicate that this should not be correlated with Late Proterozoic metadiabase dikes and flows. The chemical data (fig. 13) shows that each area of Stamford-like exposure is

chemically slightly different and therefore each area may be considered separate plutons, of 960 Ma old granite. Coarse quartz pebble conglomerate (Cdsc) very similar to that seen at Stop 1 on Day Mountain overlies the Stamford.

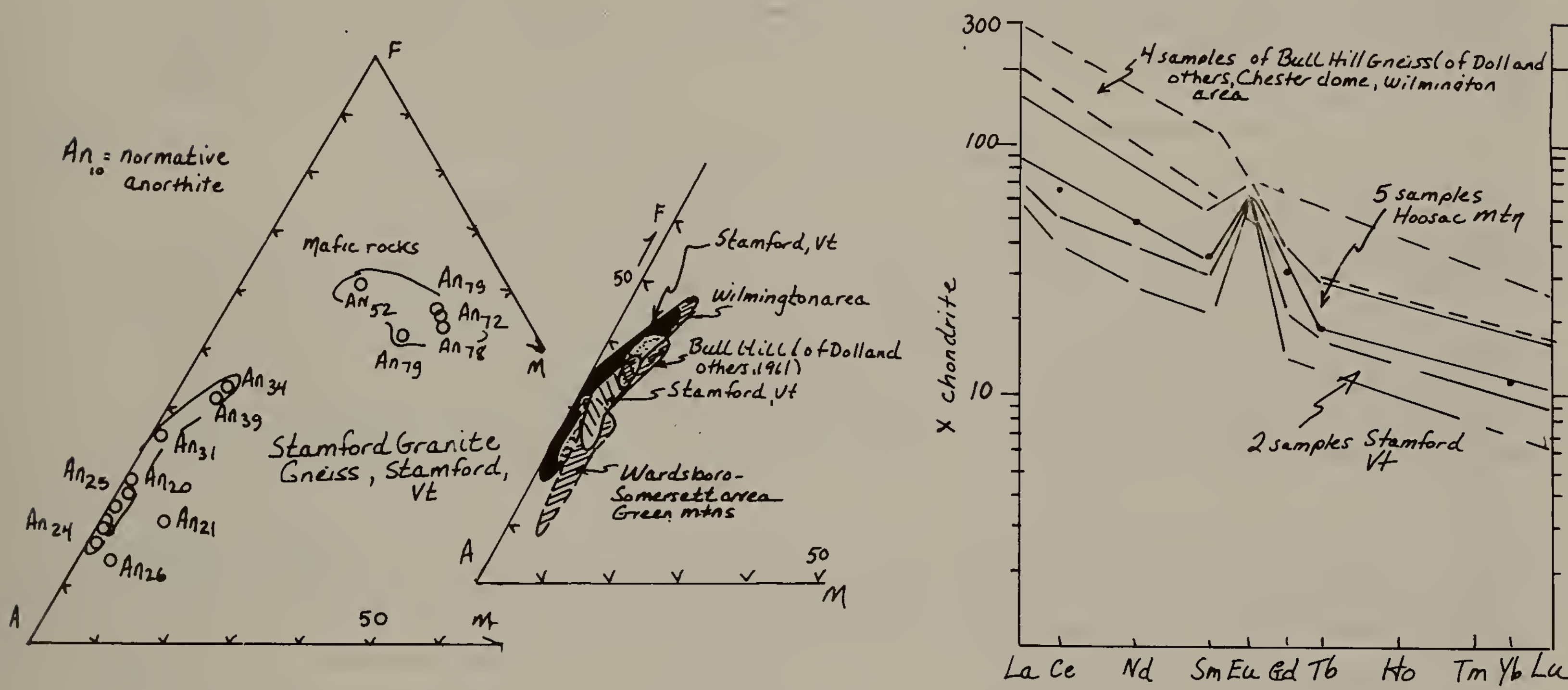


Figure 13. Some chemical characteristics of 960 Ma old post-tectonic granitoids Stamford Granite Gneiss and Bull Hill Gneiss of Doll and others , 1961, showing variable major element and trace element geochemistry of separate plutons.

We will return to Rt. 8, go south to Rt. 2 and follow Rt. 2 through North Adams and Williamstown and take Rt. 7 north toward Bennington. A partial log follows: from intersection Rt. 7 and 2, west of Williamstown, travel 7.2 miles north on route 7 to Pownal Center. At Pownal Center (brown furniture barn on right) turn right on Barber Pond Rd.,in 2.2 mi turn right on dirt road (County Rd.), 0.4 mi bear left at Y and proceed 0.9 mi uphill to park on right at sharp bend to left (total approx. 18 miles from Stop 6.

78.6 STOP 7. Lake Hancock thrust and cover sequence on the southwest flank of the Green Mountains, Pownal quadrangle.

These excellent exposures of a fault zone were first shown to Ratcliffe by Jack MacFadyen in 1959 who mapped this area (MacFadyen, 1956). Mylonitic dark schist of the Dalton Formation overlies black carbonaceous phyllites of the Middle Ordovician Walloomsac Formation. Abundant minor, fold-thrust folds, mullion structure and asymetric minor folds indicate thrusting from the southeast approximately S. 75° E. This fault can be traced northeastward where it is expressed by muscovite-chlorite rich phyllonite and mylonite gneiss in the basement gneiss. We can trace the mylonite zone associated with this fault northward to the eastern margin of the Woodford synform. Southeast plunging lineations are common on all faults in the Green Mountains and in the gneiss domes to the east and we believe these are all data from the same, late Taconian thrust fault event (fig. 9).

Retrace route to Rt. 7 via Barber Pond Rd.

82.1 Turn right on Rt. 7

88.4 Follow Rt. 7 to Rt. 9 intersection, downtown Bennington, turn right, go 3.9 miles to Woodford Hollow and park on L by white church

92.3 STOP 8. Traverse through a section of cover-sequence rocks, western margin Green Mountains massif.

Along the north side of City Stream just east of Harbor Road in Woodford Hollow is a well-exposed section of the late Precambrian to Cambrian cover-sequence rocks (Dalton Formation and Cheshire Quartzite) which overlie the gneisses at the western margin of the massif in unconformable or locally faulted contact. In this exposure bedding dips steeply eastward and, combined with observations from cross-beds and scour-and-fill features in these rocks, indicate that the section here is overturned and younging to the west. From the road eastward the section consists of (1) nonpebbley, massive, tan, vitreous quartzite with minor flaggy beds, considered to be Cheshire Quartzite (Cc); (2) massive quartzite without flaggy layers (Cc); (3) rusty-grey flaggy feldspathic quartzite of the uppermost Dalton Formation (CZd); (4) rusty, sulfidic, carbonaceous black quartz-muscovite schist (CZdb); (5) rusty flaggy quartzite (CZd); (6) rusty sulfidic black schist (CZdb); (7) more flaggy quartzite (CZd); and (8) massive tan pebbley quartzite and blue-quartz-pebble conglomerate of the basal Dalton unit (CZdc) (figs. 8 and 14). To the east, the contact with the basement (not seen in this traverse) is here mapped as a high- angle, north-south trending, east-dipping reverse fault (fig. 14) of Paleozoic (Taconian?) age, which has brought granitic gneiss of the basement over the cover-sequence rocks (for another description of the basal contact see Skehan, 1972). In many other places the contact is clearly an unconformity. The fault mapped here is responsible for producing a large-amplitude north- south-trending fold of which this section represents the overturned western limb (fig. 14). Another fault to the west in Woodford Hollow is also associated with folds of this type in the cover sequence which have steeply dipping axial surfaces. These are considered to be second- generation (F2) folds, which deform bedding and an early bedding-parallel schistocity. Possible F1 folds associated with the early schistocity are not readily visible here in outcrop but are believed here to trend east-west andcause the truncation of different cover-sequence units against the north-south trending late faults as shown diagrammatically in figure 9. The Dalton Formation, in contrast to the overlying Cheshire Quartzite, may show considerable lateral variation in composition and thickness, particularly in an east-west direction. The Woodford inlier, three kilometers to the east, contains a similar basal section of conglomerate and flaggy quartzite which however, is followed by a very thick section of black phyllitic schist, with the total thickness of the Dalton in the inlier being much greater than that of the section exposed here (figs. 8 and 14, see Skehan, 1972, for description of some of the Woodford inlier rocks). The sequence and general character of the lithologies is more uniform from north to south. In fact, strong similarities exist between the type Dalton section at Stop A-1 and rocks exposed along the western margin of the Green Mountains at least as far north as Manchester, Vt.

93.8 STOP 9. Roadcut in layered basement gneiss of the core of the Green Mountains.

This large exposure shows the strong gneissic banding and some of the compositional variation typical of the layered paragneisses mapped in the southern Green Mountains as the Harmon Hill Gneiss of Skehan (1961). The layered paragneisses have been subdivided in this study into banded biotite-rich rocks (Ybg), fine-grained hornblende amphibolites (Ya), diopside-bearing calc-silicates with minor marbles (Ycs), and rusty-ribbed garnetiferous blue vitreous quartzite (Yq) (fig. 6). Typical "Ybg" as displayed here contains alternating quartzofeldspathic and mafic (biotite plus or minus hornblende) layers on a scale of one to several centimeters. In this outcrop the general composition ranges from a quartz-poor rock with equal amounts of biotite and hornblende at the western end of the outcrop to a more felsic gneiss with only biotite as a mafic mineral at the eastern end. The quartzofeldspathic layers consist both of quartz-plagioclase- rich zones probably derived from a sedimentary protolith and quartz-plagioclase-microcline (aplite and pegmatite) lenses and stringers injected as a granitic melt, possibly derived from the large body of biotite granite gneiss (Ygg) mapped to the north and west or alternately derived migmatitic lenses (Microcline Gneiss of Skehan, 1961). The concordance of these lenses and stringers with the overall gneissosity testifies to the fact that intrusion of the granitic material and/or partial melting occurred during or prior to the major deformational episode that produced the gneissic fabric and the tight folds seen in the exposure. Samples of hornblende and biotite from this outcrop were dated using the 40Ar/39Ar technique by Samuel Mukasa (Sutter and others, 1985). Two hornblende separates yielded total-gas ages of 1127 Ma and 928 Ma, and a biotite gave a plateau age of 837 Ma (fig. 11). Thin-sections of the samples from which the dates were obtained show a fabric typical of undisturbed Middle Proterozoic rocks in which embayed interlocking grains with triple-junction boundaries occur in a flattened, granoblastic texture. The lack of textural evidence in the samples for a biotite grade Paleozoic deformational event therefore agrees well with age data. Throughout the Appalachians basement, rocks having similar relict Middle Proterozoic biotite and hornblende textures yield Proterozoic cooling ages. Paleozoic deformation is evident in this outcrop,

however, in the form of a mylonitic zone in the middle of the exposure which parallels the high-angle west-dipping gneissosity and as a faint cleavage locally crosscutting foliation at the east end of the outcrop. In these zones a secondary flattening of grains can be seen in thin section accompanied by extensive retrogradation of plagioclase to epidote and sericite and quartz ribboning. Biotite has abundant epidote inclusions and a pale red to green color, with no evidence of having been recrystallized. Although cover rocks to the west contain biotite Paleozoic temperatures were not high enough to either degas or to recrystallize biotite in the gneisses. Continue east on Rt. 9 to intersection Rt. 8 and Rt. 9 outcrops of rusty sulfidic gneiss and calc-silicate rocks

105.5 STOP 10. Taconian retrograde fabrics in Middle Proterozoic Gneiss.

Outcrops of biotite-plagioclase gneiss, and rusty sulfidic schists and calc-silicate rocks of unit Yrr are exposed to the north of the intersection. Mylonitic retrograde foliation crosscuts gneissic layering in the outcrop. In thin section lepidoblastic new biotite forms this new foliation. 40Ar/39Ar biotite plateau age of 436.5 Ma (GM4, fig. 11) indicates that this foliation and structual overprint are Taconian. Importantly Acadian overprint greater that biotite grade does not appear to have affected rocks along the eastern margin of the massif as shown by samples BM4 and BM5 to the north. A strong Paleozoic biotite grade fabric is present in ductile deformation zones in this area as shown in figure 9. Strongly foliated biotite gneiss at GM-5 to the north produced a Taconian cooling age of 446 Ma (Sutter and others, 1985).

From this point continue to east on 9 across valley marking the trace of the Searsburg thrust. Exposures of biotite gneiss at the sharp bend in the road at the Harriman Reservoir yielded the Acadian cooling age of 363 Ma (GM29) figure 11, suggesting that the transition between Acadian overprinted zones and Taconic zones is abrupt and may be affected by Acadian faulting.

113.6 continue east to intersection of Rt. 9 and Rt. 100, downtown Wilmington. Turn north on Rt. 100 and proceed north 1.7 miles to Elementary School on west side of Rd.

115.3 STOP 11. Hornblende amphibolite and gneiss in Wilmington antiform and discussion of Ar/Ar results.

Hornblende and biotite from this outcrop (GM28, fig. 11) yielded 40Ar/39Ar ages of 376 and 377 Ma respectively. Although originally mapped as the Turkey Mountain Member of the Hoosac Formation by Skehan (1961) this rock may actually be basement gneiss. In this general area strong subhorizontal mylonitic foliation is prevalent in rocks above the Wilmington fault. Crosscutting pink pegmatoids as veins and pods are common. These granitic pods are non-deformed and suggest that post-thrusting high grade metamorphism affected these gneisses. The Ar/Ar data suggest that hornblende and biotite passed through their effective blocking temperatures nearly simultaneously, thus suggesting quick cooling from an Acadian high temperature event. Textures and mineralogy in nearby rocks suggest widespread post-foliation recrystallization consistent with Acadian thermal overprinting of Taconian fabric.

Fieldtrip ends - return south on 100 Wilmington, take Rt. 9 east to Brattleboro and 5 and 9 north to cross Connecticut River. Follow Rt. 9 to Keene.

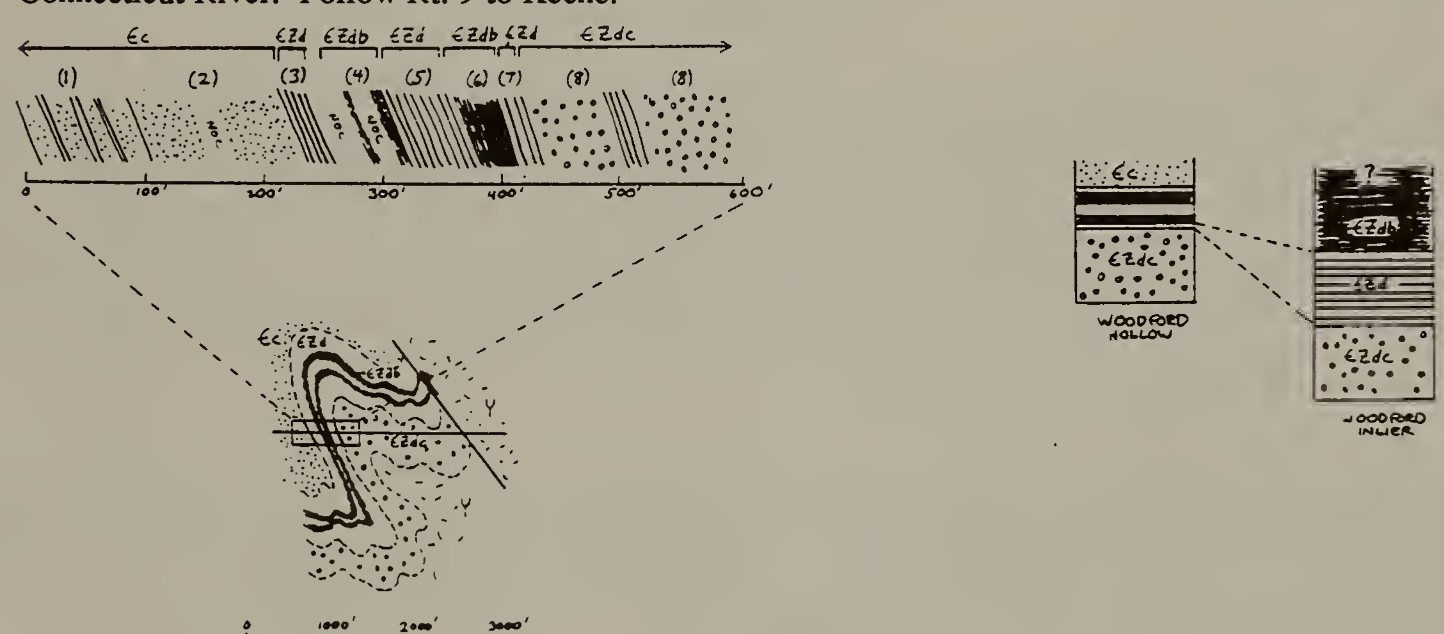

Figure 14. Cross section through Cheshire Quartzite and Dalton Formation on east side of Woodford Hollow cover sequence, Stop A-8 and comparison of cover sequence in Woodford Hollow and Woodford synform.

# STRATIGRAPHY AND STRUCTURE OF THE FALL MOUNTAIN AND SKITCHEWAUG NAPPES, SOUTHWESTERN NEW HAMPSHIRE

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## INTRODUCTION

On this field trip, we present results of field research on the Skitchewaug and Fall Mountain Nappes. These nappe structures are exposed in two adjacent Acadian structural terranes, the Bronson Hill Anticlinorium and the Merrimack Synclinorium (figure 1). Our research on the nappes was directed towards understanding the tectonic relationships between these two large structural terranes.

Various geological studies have shown that the sedimentalogical, structural, and metamorphic histories of these two terranes are significantly different (Lyons, 1979; Thompson et al., 1968; Robinson and Hall, 1980). Rocks in the Bronson Hill Anticlinorium consist of Ordovician volcanics and sediments, Silurian shallow water marine sediments, and Devonian turbidites folded into a series of domed nappes (Thompson et al., 1968). Whereas, the rocks in the Merrimack Synclinorium consist of Ordovician sediments without volcanics, deep water Silurian sediments, and Devonian turbidites folded into nappes showing several stages of later folding (Lyons, 1979; Hatch et al., 1983). In addition, metamorphism in the Bronson Hill Anticlinorium is generally staurolite grade (ranging from garnet to sillimanite grade) and occured early in the deformation history (Thompson et al., 1968). This metamorphic style contrasts markedly with that observed in the Merrimack Synclinorium, where peak metamorphism is higher grade (sillimanite to sillimanite-potassium feldspar-cordierite grade) and reached peak conditions later in the deformation history (Chamberlain, 1985). Accounting for these sedimentalogical, structural, and metamorphic differences is an important problem facing the geologist interested in the tectonic evolution of the Acadian Appalachians

Toward this aim we have studied the statigraphy, structure, and metamorphism of the Fall Mountain and Skitchewaug Nappes in exposures that are found both on the east and west flank of the Bronson Hill Anticlinorium in southwestern New Hampshire. Our results are consistent with a tectonic history that involves: 1) deposition of sediments in an east-thickening basin during the Silurian; and 2) subsequent emplacement of this east-thickening package of sediments onto the more shallow water shelf sediments during the Devonian as fold and thrust nappes.

## FIELD TRIP AREA

Our field trip concerns the metasediments in the Walpole, Bellows Falls, Lovewell Mountain, and Stoddard 7 1/2 minute quadrangles. The metasediments in this region have been the subject of study for over a century, and earlier interpretations are given in a large number of published geologic reports (Kruger, 1946; Moore, 1949; Heald, 1950; Chapman, 1952; Thompson et al., 1968; Dean, 1976).

## STRATIGRAPHY

In the most recent published paper on the stratigraphy and structure of this region Thompson et al., (1968) assigned the rocks to the classic Bronson Hill stratigrphy defined by Billings in 1937. According to these authors interpretation the rocks in southwestern New Hampshire consisted of the following sequence, from bottom to top: Ordovician mixed felsic and mafic volcanics (Ammoonosuc Volcanics), Ordovician rusty weathered schists (Partridge Formation), a Silurian conglomerate (Clough Formation), Silurian calc-silicate (Fitch Formation), and a Devonian gray weathered turbidite (Littleton Formation). More recent mapping in northern Maine (Moench and Boudette, 1970; Boone et al., 1970) and New Hampshire (Hatch et al., 1983) have shown that the stratigraphy in both the Bronson Hill Anticlinorium and Merrimack Synclinorium is more complex and different than the stratigraphy originally defined by Billings (1937).

Our remapping of the metasediments exposed in the Skitchewaug and Fall Mountain nappes suggests that the stratigraphic sequence of Billings (1937) cannot be directly applied to these rocks. Our results suggest that rocks in these two nappes show many similarities to some recently defined stratigraphic sequences in Maine (Moench and Boudette, 1970) and central New Hampshire (Hatch et al., 1983). Moreover, the stratigraphy presented here gives direct evidence that the Silurian stratigraphy becomes increasingly complex eastward from the main axis of the Bronson Hill Anticlinorium.

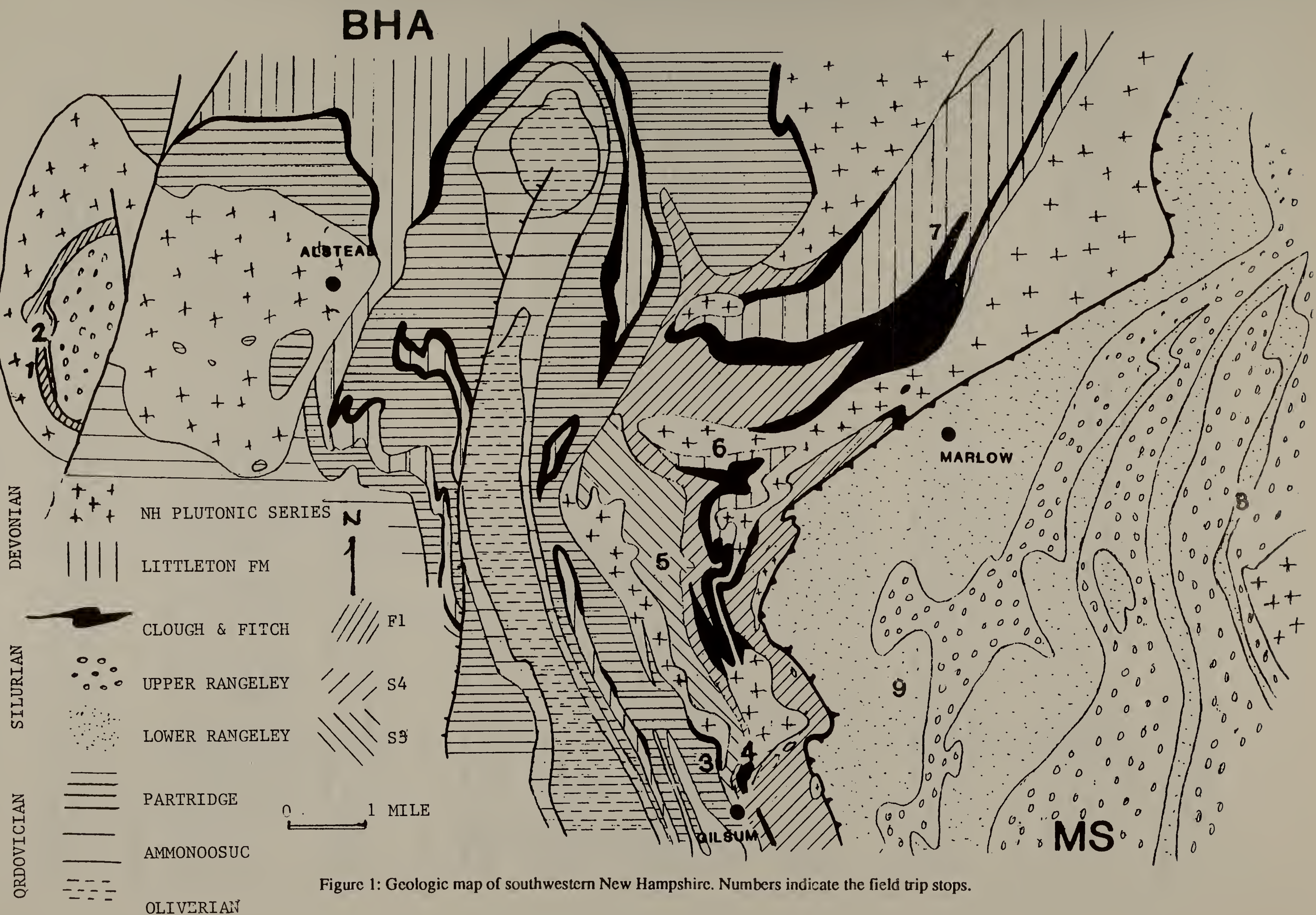



Figure 1: Geologic map of southwestern New Hampshire. Numbers indicate the field trip stops.

We subdivided the rocks in the Fall Mountain and Skitchewaug nappes into 12 lithologically distinct units. The 12 units occur in the Skitchewaug and Fall Mountain nappes, and the stratigraphy of these two nappes are somewhat different (figures 1 and 2).

The relative ages for the stratigraphic columns presented in figure 2 were determined by examining stratigraphic topping directions in rocks here assigned to the Littleton Formation, and by observing consistent lithic sequences throughout the field area. We point out that we found no recognizable fossils in any of the units in this area so long range correlations are made solely on the basis of lithologic and sequential similarities to stratigraphic sequences observed in other regions of the orogen.

*Skitchewaug nappe:*

The stratigraphy presented here is for rocks belonging to what we believe to be the upper limb of the Skitchewaug nappe exposed on the east-side of the axis of the Bronson Hill Anticlinorium. The details of this stratigraphic sequence are given in Chamberlain (1985). We do not present the stratigraphy of the Skitchewaug nappe exposed to the west of the Bronson Hill Anticlinorium because it has been presented in a previous paper (Thomspon et al., 1968)

East of the Alstead and Keene gneiss domes we recognize 8 stratigraphic units within the Skitchewaug nappe. The lowest unit (Unit S1) consists of a massive, rusty weathered schist. These schists frequently contain abundant lenses of pink coticule and amphibolite layers. Stratigraphicaly above this units is a thin (0 to 50 meters), polymict, quartz-matrix conglomerate and plagioclase-biotite granulite (Unit S2). The conglomerate is overlain in many places by a few meters of thinly bedded calc-granulite.

Above Units 1 and 2 is a thick (greater than 1000 meters), gray to blood-red weathered, massive, sillimanite-bearing schist (Unit S3). These schists contain pod-shaped calc-silicate clasts, which are diagnostic of the unit. The schists of Unit 3 grade into a massive (750 meters), gray weathered, sillimanite-bearing schist that contains abundant lenses of conglomerates and plagioclase-biotite granulites (Unit S4). Both the conglomerates and the granulites within this unit are generally thin (1 to 5 meters thick) and are laterally discontinuous. Two types of conglomerates are present: polymict, schist matrix conglomerates; and polymict, granulite matrix conglomerates.

Above these rocks is a conglomerate (Unit S5) that is laterally continuous over much of the field area. This conglomerate has predominantly a quartz-matrix and contains clasts of mostly vein quartz, with minor clasts of pelite, amphibolite, and quartzite. Overlying the conglomerate is a distinctive calc-silicate unit (Unit S6). This unit consists of rusty weathered calc-silicates at its base, and grades into gray weathered bedded calc-silicates at its top.

Stratigraphically above the calc-silicates of Unit S6 is a thick (greater than 1000 meters) of graded bedded schists. Topping direction of this unit (Unit S7) are easy to determine because compositional gradations between sandy and aluminous portions of each bed is gradual.

*Fall Mountain Nappe:*

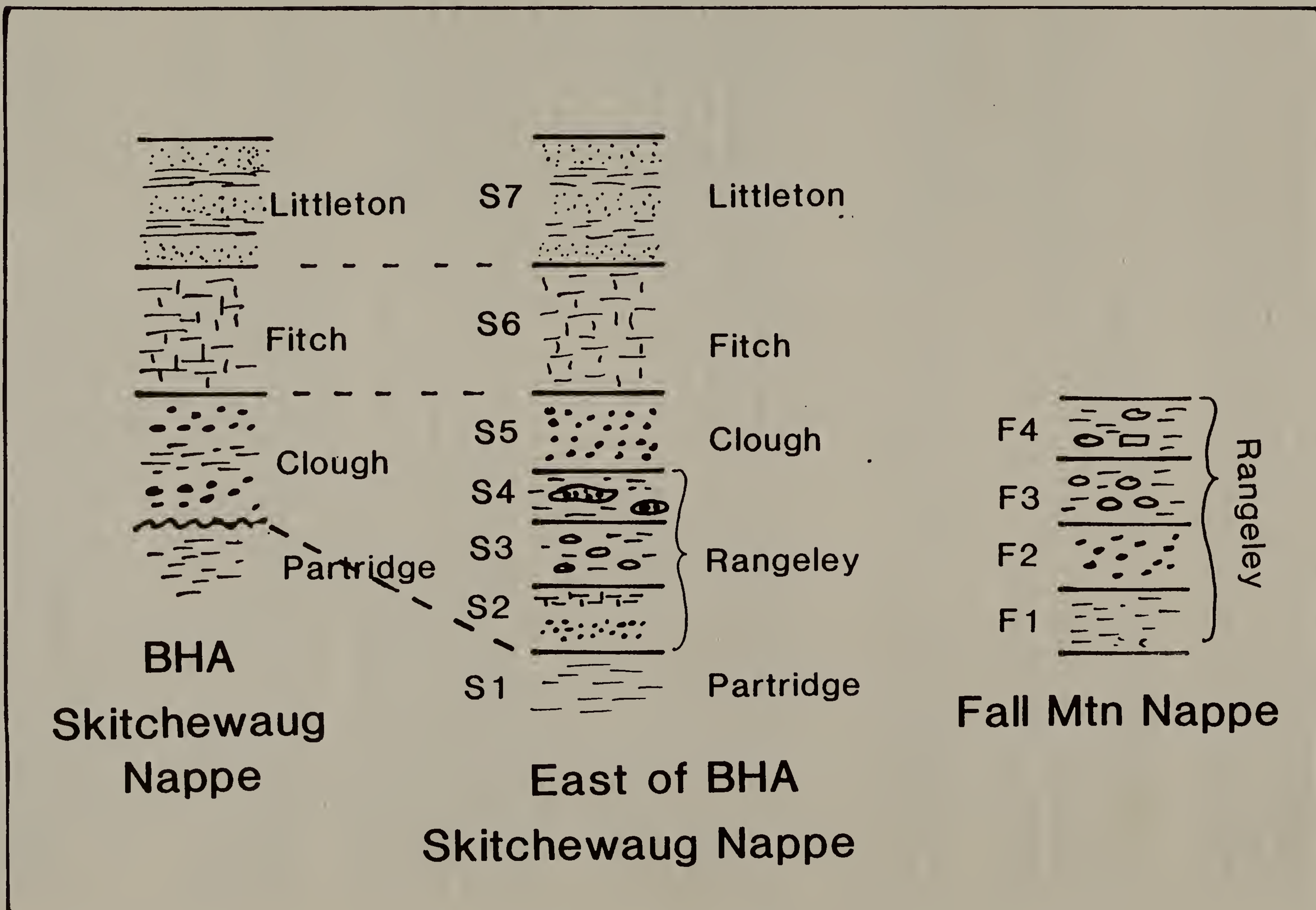
The stratigraphic sequence observed in the structurally higher Fall Mountain Nappe is different than the stratigraphy for the Skitchewaug Nappe. Four stratigraphic units are present in the Fall Mountain Nappe (Chamberlain, 1985, Allen, 1985). Topping directions of these four units have been impossible to determine. We assigned their relative ages based on lithic similarities to rocks in the Merrimack Synclinourium.

The lower most unit (F1) is generally a massive, (though sometimes bedded), gray weathered, sillimanite-bearing pelite. This unit typically has large sillimanite crystals that have the appearance of "turkey tracks". Above this unit, is a thin (less than 30 meters) polymict, schist-matrix conglomerate (F2). This conglomerates appears as lenses between Units F1 and F3. Overlying the conglomerate is a thick (>1000 m), massive, blood-red weathered schist that contains calc-silicate pods (F3). This unit is identical to Unit S3 in the Skitchewaug Nappe. At the top of the sequence, is a massive, sometimes bedded, rusty weathered schist (F4). The schists of unit F4 contain abundant calc-silicate beds and calc-silicate pods.

*Correlations:*

We have correlated the stratigraphic sequence presented here with stratigraphies in the Bronson Hill Anticlinorium (Billings, 1937), the Merrimack Synclinorium in central New Hampshire (Hatch et al., 1983), and rocks between the main axes of these two terranes in northern Maine (Moench and Boudette, 1970). All of the

Figure 2: Straigraphic columns for the Skitchewaug Nappe on the Bronson Hill Anticlinorium (BHA); the Skitchewaug Nappe east of the BHA; and the Fall Mountain Nappe. Schists are shown as dashes; conglomerates as large dots; calc-silicates and granulites as a brickwork pattern; turbidites as alternating small dots and dashes; open circles enclosed within dashes represent pods within massive schist; lenses of black dots within dashed pattern represents conglomerate lenses in a schist matrix. The abbreviations S1, F1, etc. are described in the text.



correlations that we propose are based on lithic and sequential similarities between the units discussed above and rocks in these three other regions.

In our interpretation, the rocks immediatley east of the Alstead and Keene domes are Ordovician through Devonian strata; and these strata belong to the upper limb of the Skitchewaug Nappe. We suggest the following correlations: Unit S1=Partridge Formation (Ordovician); Unit S2 Lower Rangeley (Silurian), Unit S3= Middle Rangeley (Silurian), Unit S4 = Upper Rangeley (Silurian); Unit S5= Clough Formation (Silurian); Unit S6=Fitch Formation (Silurian); and Unit S7=Littleton Formation (Devonian).

The stratigraphy of the Skitchewaug nappe is different from the stratigraphies of the Bronson Hill Anticlinorium and the Merrimack Synclinorium, and is most similar to rocks in a similar structural setting in northern Maine (Moench and Boudette, 1970). In the Bronson Hill Anticlinorium, the Partridge Formation is overlain by the conglomerates of the Clough Formation. However, on the east flank of the Bronson Hill terrane a thick sequence of mixed schists and conglomerates (S2-S4) lie between rocks assigned to the Partridge and Clough Formations. A similar thick sequence of Silurian rocks are also present above the Ordovician sedimentary rocks in Maine (Moench and Boudette, 1970).

Rocks in the Fall Mountain Nappe (F1-F4) we believe to be correlative with the rocks assigned to the Rangeley Formation in central New Hampshire (Hatch et al., 1983). Our F1 and F2 is unlike any units observed in central New Hampshire, but are similar to Rangeley rocks found in Maine. Our units F3 and F4 we correlate with the Lower Rangeley and Upper Rangeley, respectively.

## STRUCTURE

Earlier mapping along the Bronson Hill Anticlinorium (Thompson et al., 1968) suggested that the rocks belonged to several major refolded nappes. Our research confirms this earlier structural interpretation. However, we suggest a slightly more complex structural history that involves three phases of folding and thrusting. The earliest period of deformation produced west-vergent fold and thrust floored nappes and the dominant foliation in this area. The nappes were later folded about northeast-southwest trending, steeply plunging axes. The second stage of folding created a spaced crenulation cleavage and abundant tight to isoclinal minor folds. The latest period of deformation produced open folds with shallow-plunging, east-west axes. There is no recognizable cleavage asociated with the last folding event.

The tectonic break between the Skitchewaug and Fall Mountain Nappes, where these structures are exposed east of the gneiss domes, is a thrust fault. The fault is marked by a stringer of Kinsman Quartz Monzonite (called the Huntley Mountain Spur) that extends off the west side of the Cardigan pluton. The fault contiunues south into the Monadnock quadrangle, where it has been named the Chesham Pond Thrust (P. Thompson, 1984). In the Monadnaock area, the thrust trends to the east-west across the northern part of the quadrangle. The thrust is also exposed between the Fall Mountain and Skitchewaug Nappes west of the gneiss domes in Bellows Falls, Vermont. In this region, the fault is thought to occur within the Bethlehem Gneiss of the Bellows Falls pluton (Allen, 1985). It is interesting to note that the Bellows Fall pluton and the igneous rocks in the Huntley Mountain Spur are similar both in their structural position and lithology.

Two lines of evidence support the interpretation that the boundary between the Skitchewaug and Fall Mountain nappes is a major thrust fault. First, the stratigraphies west of the Huntley Mountain Spur is more similar to rocks exposed on the Bronson Hill Anticlinorium, whereas the rocks east of the Spur are more similar to rocks observed in central New Hampshire. The juxtaposition of vastly different stratigraphic succesion suggests major westward transport of the rocks assigned to the Fall Mountain Nappe. Second, mapping along the tectonic break show that stratagraphic units pinch out along the contact between the Skitchewaug and Fall Mountain Nappes.

## SUMMARY

Our interpretation of the stratigraphy and structure of southwestern New Hampshire is consistent with an Acadian history that involves continued closure of an east-thickening Silurian basin during the Devonian. Along the east side of the Bronson Hill Anticlinorium, we recognize a thick sequence of schists and interbedded conglomerates that occur between the Clough Formation and the Ordovician schists of the Partridge Formation. These intervening schists and conglomerates, in our interpretation, represent an east-thickening transitional facies correlative to the Silurian shelf sediments (Clough) of the Bronson Hill and basin sediments (Rangeley) of the Merrimack Synclinorium. Our interpretation is consistent with an earlier hypothesis that during the Silurian an east-thickening basin existed and the Bronson Hill Anticlinorium was a major topographic high (Boone et al., 1970).

During the Acadian the deep water sediments were thrust over the shallow water sediments of the Bronson Hill Anticlinorium. Thus, succesively higher nappes have stratigraphies that contain less conglomerates and more deep water schists. This is best observed when comparing the startigraphic succesions in the Skitchewaug and Fall Mountain Nappes.

ACKNOWLEDGEMENTS

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ROAD LOG

Assembly point: Keene State College Commons Parking Lot: 8 am, Friday, Oct. 14:
This trip will involve several strenuous hikes and much travel on steep, narrow gravel roads. The mileage log begins from the first stop. Topographic Maps: either the old Bellows Falls (NH and VT) and Lovewell Mountain (NH) 15' Quadrangles or the new Bellows Falls, Walpole, Lovewell Mountain, and Stoddard 7 1/2' quadrangles.

From Keene take NH Route 12 north to the first stop at the Vilas bridge, connecting North Walpole, NH with Bellows Falls Vermont, about 20 miles.

0.0 **STOP 1:** Park on the east side of the road, just north of the Green Mountain Railroad's yard. Here we will be looking at turkey track sillimanites of unit F1 in contact with the Bellows Falls Pluton.

0.4 North on RT. 12 to Milt's Corner Market, bear right.

0.6 Turn sharp right up hill, past school.

0.7 Turn left, uphill.

1.0 STOP 2: Park at end of road to begin hike up Fall Mountain. Keep vehicles close together as parking is limited. This hike will be about 3 miles roundtrip and will involve over 600 feet of elevation gain. We will see all units of the Fall Mountain Nappe stratigraphy on this hike.

1.6 Return to RT 12, travelling south back past Vilas bridge.

3.1 Junction with RT 123, turn left following sign for Alstead.

3.3 Bear left again on Cold River Road.

OPTIONAL STOPS: On the Cold River Road there are several optional stops we can make if time and interest permit, including Partridge Formation underlying the Bellows Falls Pluton, and as xenoliths within the Pluton.

7.1 Intersection of the Cold River Road and RT 123. Continue straight across, now travelling "south" on RT 123.

9.0 Junction with RT 12-A. Turn right then left staying on RTs 123 and 12-A through Alstead village. Stay on RT 123 past junctions with 123-A to Acworth and 12-A to Keene. You will be driving over rocks of the Skitchewauge Nappe and Alstead Dome.

14.5 East Alstead, turn right at blinking light, following sign to Gilsum.

19.4 STOP 3: At the top of hill, park on right side as far over as possible. Here we will be looking at the contact between units S1(Partridge) and S2 (Rangeley) in the upper limb of the Skitchewaug Nappe.

20.1 Near bottom of hill,as we are approaching Gilsum village, turn abrupt hard left onto narrow road. Follow paved road until pavement ends.

20.5 STOP 4: Park in turn out to the left; avoid blocking the road. From this parking area we will see several different outcrops of units S3, and S4.

20.9 Turn right, downhill, onto paved road.

21.3 Turn right onto RT 10, into Gilsum.

21.6 Turn right into Gilsum Village

21.7 LUNCH STOP: Turn right and park in front of meeting house lawn. We will stop here for lunch, after which we will continue north out of Gilsum village, returning to East Alstead.

25.0 STOP 5: park on right side as far over as possible. Here we will see large amphibolite pods within the Rangeley Schists belonging to Unit S3.

26.3 Continue on the right hand fork, straight ahead.

27.4 East Alstead again, turn right on RT 123, towards Marlow.

28.2 Turn right at some mailboxes just past a barn onto Roger's Road

29.2 STOP 6: Fork in the road, parking and turn around will be difficult, but park on either side of the road, keeping the road clear. From this parking area we will again disembark to see several outcrops of Unit S5 (Clough-the upper most conglomerate) and S7 (Littleton).

29.5 Ahead on left fork, turn arounds are available in several driveways.

30.6 Turn right on RT 123 again headed toward Marlow.

31.9 OPTIONAL STOP: Schist matrix conglomerate of Unit S4 (Rangeley).

34.1 Entering Marlow village, turn left

34.4 Join RT 10 North

36.6 **STOP 7:** Gee Mill, park on right side as far over as possible, past turn for Sand Pond road. Here we will see graded bedded Littleton Fm (Unit S7), and Fitch Fm (Uniy S6) in the hinge of a large F3 fold.

Turn around, heading south on RT 10, back through Marlow.

40.5 Junction with RT 123 East, turn left following sign for Stoddard.

43.6 **STOP 8:** Turn left into Pitcher Mountain Parking Lot. From here we will take a short hike up Pitcher Mountain. At the summit we will see Units F3 and F4 (assigned to the Rangeley Fm of central NH) that are in the Fall Mountain Nappe.

Turn around, heading west on RT 123, back towards RT 10.

46.6 At sharp bend in road, turn left onto short road that will connect with RT 10.

46.9 Turn left onto RT 10 South.

49.2 **STOP 9:** Park on right side as far over as possible. Spectacular outcrops of high-grade Rangeley Fm. (Unit F3) of the Fall Mountain Nappe.
Continuing south on RT 10 will bring you to Gilsum and then back to Keene, about 10 miles.

# STRATIGRAPHY, STRUCTURE, AND METAMORPHISM OF THE 'DORSAL ZONE', CENTRAL NEW HAMPSHIRE

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## INTRODUCTION

The purpose of this field trip is to examine the bedrock geology of the Gilmanton, New Hampshire, 15 minute quadrangle and immediate surrounding area. The stratigraphy, which forms a link between the sections extending east from the Bronson Hill anticlinorium and west from the Campbell Hill Nonesuch River fault zone, and a new structural model for the complexly deformed metasedimentary rocks of this area will be presented. The nature of regional metamorphism and radiometric age dates that constrain the timing of plutonism and metamorphism will be examined. The ultimate goal of this field trip is to present a coherent geologic model for the region, based on structural, petrologic, and geochronologic data.

## REGIONAL GEOLOGIC SETTING

This study focuses on a group of rocks belonging to the Kearsarge-Central Maine synclinorium (Lyons and others, 1982) formerly the Merrimack synclinorium of Billings (1956) and also called the Merrimack Belt by geologists in Massachusetts (Zen et. al. 1983). This is interpreted as a large Silurian to early Devonian depositional basin that was multiply deformed and metamorphosed during the Acadian and Alleghanian (?) orogenies.

Since the introduction of the name Kearsarge-Central Maine synclinorium by Lyons and others in 1982, there has been semantic confusion when describing the major geologic features in this region. The recent mapping by students of John Lyons at Dartmouth College, Peter Robinson at University of Massachusetts, and Wallace A. Bothner at the University of New Hampshire has shown that there is not just one synclinorium in this tract of rocks, but several, with intervening anticlinoria (for example see; E. Duke, 1984; P. Thompson; 1985, Eusden, 1984 and 1988). It is clear that the Kearsarge-Central Maine synclinorium is an inadequate descriptive term for the entire region.

Eusden and others (1987) and John Lyons (this volume) now recognize the following major structures in the Kearsarge-Central Maine synclinorium of New Hampshire; the Kearsarge-Central Maine synclinorium (KCMS) proper; the Lebanon antiformal syncline (LAS); the Central New Hampshire anticlinorium (CNHA); the Boundary Mountains anticlinorium; and the Chocorua syncline. For this report the term Central Maine Terrane (CMT) (Zen et. al., 1986) will be used to include the rocks that extend northeasterly from Connecticut to Maine and from the Bronson Hill anticlinorium east to the Campbell Hill-Nonesuch River fault zone (CHNRFZ) (Figure 1). Reference to the KCMS is restricted to axial trace of the synclinorium proper through the Devonian Littleton Formation. This reconcilliation of regional nomenclature will probably not sate geologists that feel the belt should extend southeast beyond the CHNRFZ to include the formations of the Merrimack Group (Berwick, Eliot and Kittery Formations), and hence would prefer to keep the term Merrimack Belt or synclinorium as originally described by Billings (1956). However, many geologists, myself included, believe that the aforementioned Merrimack Group and Massabesic Gneiss Complex is part of a Precambrian exotic terrane quite different in origin from the rocks of the CMT (Bothner and others, 1984; Lyons and others, 1982 and 1986; Gaudette and others, 1984; Naylor, 1985).

## PREVIOUS WORK

Mapping within the western part of the CMT, along the axis of the KCMS, from west-central New Hampshire to a fossiliferous Silurian and Lower Devonian section exposed near Rangeley, Maine has revised the stratigraphic interpretation of Billings (1956) (Hatch and others, 1983; Moench and Boudette, 1970; Moench, 1984). Much of what was assigned to the Lower Devonian Littleton Formation is now interpreted as a thick section of Silurian turbidites correlative to the Rangeley, Maine section (Hatch and others, 1983; Nielson, 1981; Thompson, 1983, 1984, 1985; Chamberlain, 1984; G. Duke, 1984 and E. Duke, 1984).

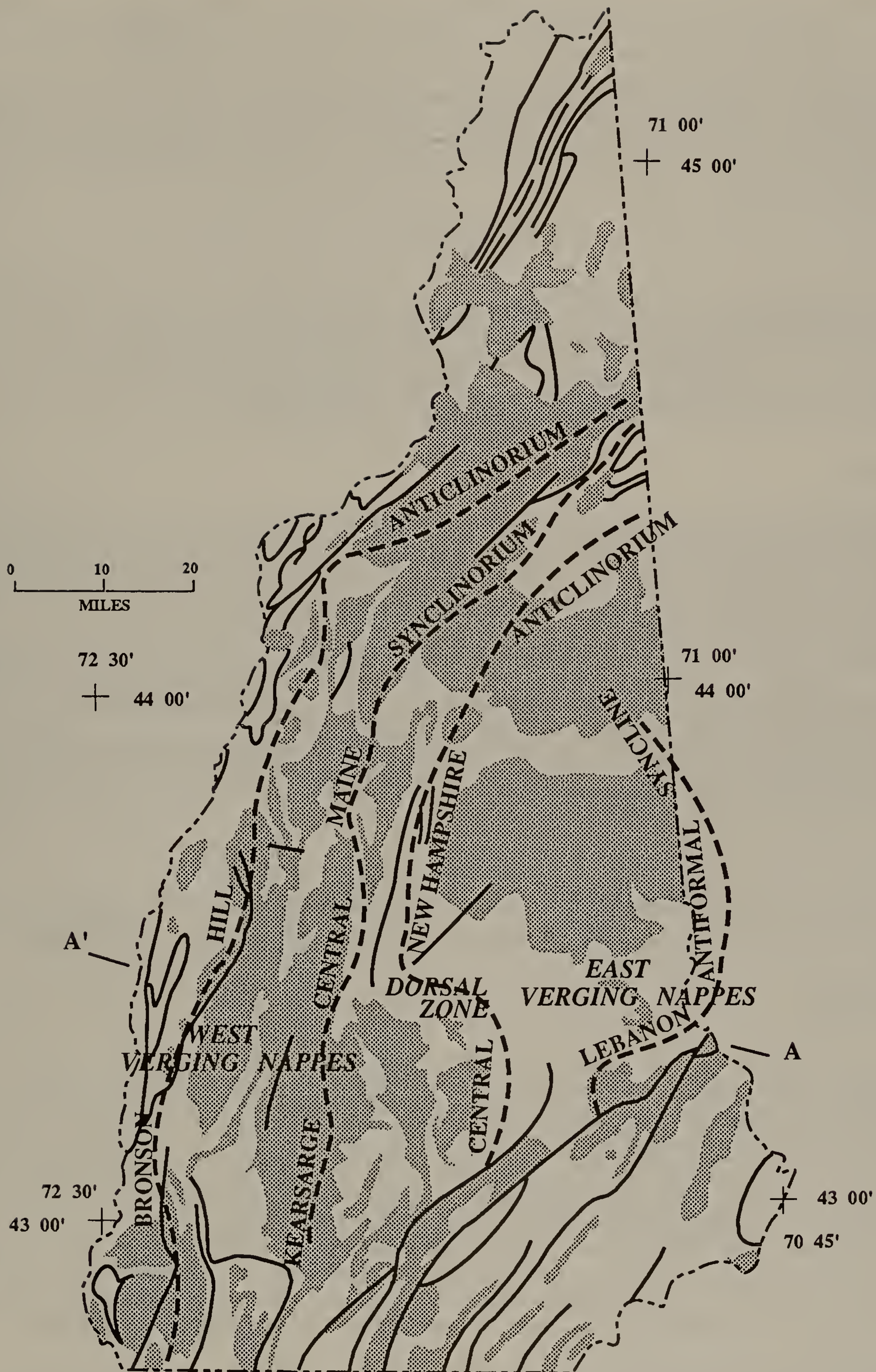



**Figure 1. Major fold structures in the Central Maine Terrane. Igneous rocks shaded, faults shown as solid black lines, axial traces of folds shown as bold dashed lines. Vergence of nappes and "Dorsal Zone" also shown.**

In southeastern New Hampshire and adjacent southwestern Maine the metasedimentary rocks in this part of the CMT were previously mapped as Lower Devonian Littleton Formation by Billings (1956), Stewart (1961, Alton quadrangle), Heald (1955, Gilmanton quadrangle), and Carnein (1976, Suncook quadrangle) and as the Shapleigh Group by Hussey (1962 and 1968, York County, Maine) and Gilman (1977 and 1978, Newfield and Kezar Falls, Maine quadrangles). Eusden and others (1984 and 1987) subdivided these rocks into five lithostratigraphic units and remapped a second major synclinorium in the Central Maine Terrane, the Lebanon Anitformal syncline. The units in this area were correlated to the Siluro-Devonian Rangeley, Maine section and the central New Hampshire section in the KCMS.

## STRATIGRAPHY

The stratigraphic section in the Gilmanton quadrangle is different from the members of the Littleton Formation as originally defined by Heald (1955). Heald recognized the Durgin Brook, Pittsfield and Jenness Pond Members of the Littleton Formation. Few similarities exist between Heald's stratigraphy and that presented here and it is proposed that the former be abandonned. The revised stratigraphy, presented below, was determined by examining primary stratigraphic topping directions at the contacts between units and, once established in critical localities, by using structural data to determine the superposition of the formations.

### Littleton Formation

In the Gilmanton quadrangle the Littleton is a coarse-grained grit or conglomerate with gray and rusty-weathering lithic fragments of the Smalls Falls Formation up to 10 cm in length, as well as angular to subrounded vitreous quartz clasts. The matrix is composed of muscovite, quartz and biotite. Eusden and others (1987) informally named the grit exposures the Wild Goose Grits and correlated them to a similiar lithology at the same stratigraphic position in the Alton and Berwick quadrangles mapped by Hussey (1962) as the Towow Grits.

The Littleton in the Alton and Berwick quadrangles occupies the axial trace of the Lebanon antiformal syncline and is the typical gray well-bedded pelitic turbidite seen throughout New Hampshire and Maine. The Wild Goose Grits are restricted to the basal part of the formation at or near the contact with the Smalls Falls Formation. Note that here the Madrid Formation is missing, implying an unconformity.

The Littleton is at the top of the Paleozoic section and because of erosion only minimum estimates of the thickness can be made. The Littleton may have been up to a kilometer or two thick prior to erosion. The Wild Goose Grits are thin, only about 100 to 300 meters thick.

### Madrid Formation

No outcrops of the Madrid Formation were found in the Gilmanton quadrangle. The formation which should be found between the Smalls Falls and Littleton Formations is missing, and has either been eroded away or was never deposited.

### Smalls Falls Formation

The Smalls Falls is a red-brown to dark brown-black, deeply rusty-weathering schist with pyrrhotite and occasional graphite. Infrequent sulfidic quartzite layers, up to several cm in thickness, are seen within the schist. The overall appearance is a well foliated, very crumbly rusty schist that is almost always crenulated by late small-scale kink-type folds. The outcrops weather easily, and as a result natural outcrops are not as abundant as those from other more resistant formations. However, the pyrrhotite renders the formation moderately magnetic and positive magnetic anomalies coincide with the map pattern in this area (Bothner and others, 1988) allowing it to be tracked easily. The upper and lower contacts are marked by an abrupt loss of the distinctive rusty weathering.

In the Gilmanton quadrangle the Smalls Falls is between 100 and 500 m thick. To the northeast in the Alton quadrangle and Berwick quadrangle it has a wide range of thicknesses. In places it is absent due to non-deposition, erosion or faulting (?). Clasts of Smalls Falls in the Wild Goose Grits suggest that the true thickness was probably greater and that erosion probably played a greater role in its thickness distribution then non-deposition.

Perry Mountain Formation

The Perry Mountain Formation is a well-bedded, gray schist and quartzite. The quartzite can be quite thick, commonly 10 cm and up to 1 m, and is almost always 'clean', meaning with little mica in it. In a few places the Perry Mountain is moderately migmatitic but it is in general not as easliy migmatized as the Rangeley Formation. In the Gilmanton quadrangle the formation has been divided into two members. The style of bedding is identical in each member. The upper Perry Mountain is characterized by pink garnet plus quartz coticules that occur within and toward the base of the quartzite beds, and also by the lack of or extremely rare calc-silicate boudins or 'footballs'. The coticules are 1 to 5 cm thick, discontinuous stringers and pods, often more complexly folded and deformed then the surrounding quartzite.

The lower Perry Mountain is characterized by calc-silicate boudins that occur most often within the quartzite beds, and by the lack of any coticules. The calc-silicate boudins have rims of biotite, quartz and plagioclase and are dark gray in color; the cores are lighter in color, white to gray, and commonly stand out as elliptical resistant knobs. They are composed of quartz, plagioclase, grossular and less frequently diopside.
The Perry Mountain is about 400 to 500 m thick. The upper Perry Mountain is about 200 to 300 m thick and the lower Perry Mountain is about 100 to 300 m thick.

Rangeley Formation

The Rangeley Formation is the most varied of all the formations in the stratigraphy. The three fold subdivision used in this study is not used elsewhere. For example, Lyons and others (1986) subdivide the Rangeley into upper and lower members only. Moench and Boudette (1970) have identified Rangeley A, B and C; though threefold in subdivision, these lithologies are coarse conglomerates and turbidites quite unlike those mapped in the Gilmanton quadrangle or elsewhere in central New Hampshire. The middle and lower Rangeley of the Gilmanton quadrangle are together equivalent to the lower Rangeley of Lyons and others (1986) and, the upper Rangeley is equivalent to their upper Rangeley.

The upper Rangeley is a red-brown, rusty weathering, often graphitic, schist and sulfidic quartzite. Unlike Lyons and others (1986) upper Rangeley there are no, or only rare, calc-silicate boudins in this unit. The upper Rangeley differs from the Smalls Falls formation by the greater abundance of quartzite beds and the much lower abundance of pyrrhotite. The quartzite beds are thin, from 1 to 5 cm. Positive magnetic anomalies, commonly associated with the Smalls Falls, are not seen in conjunction with the mapped pattern of the upper Rangeley (Bothner et al., 1988). The rock weathers easily, and is often crumbly and not well exposed; it commonly crops out in the low-lying areas of the quadrangle. The contact with the overlying Perry Mountain Formation is abrupt and marked by the disappearance of well-bedded turbidite in the Perry Mountain, and the appearance of rusty-weathering schist in the Rangeley.

The middle Rangeley is a well-bedded schist and quartzite with no calc-silicate boudins. Bedding is about 3 to 10 cm thick and graded beds are not usually seen due to the abrupt transition between quartzite and schist. This rock is quite similiar to the parts of the Perry Mountain that have no calc-silicate boudins or garnet coticules. The contact with the upper Rangeley is abrupt, and like the contact between the Perry Mountain and upper Rangeley, it is marked by the loss of rusty-weathering schist and the appearance of gray well-bedded turbidites.

The lower Rangeley is a massive to well-bedded, calc-silicate boudin-bearing, biotite granofels. It is purple to gray in color. Bedding is defined by alternating layers of schist and granofels and is highly variable in thickness, ranging from 1 cm up to 1 m. The calc-silicate boudins are elliptical and mineralogically zoned. They have dark gray rims of biotite, quartz and plagioclase and light gray to white cores of plagioclase and quartz, with green and red spots of grossular and diopside. The cores often stand out as resistant knobs in outcrop. The long axes of the boudins are parallel to the plane of bedding. Almost every outcrop in the belt of lower Rangeley has at least one boudin in it. The contact with the middle Rangeley is gradational and was drawn on the first appearance of calc-silicate boudins as one moves down-section.

Rangeley Formation Migmatites

Much of the Rangeley Formation is extensively migmatized in the Gilmanton quadrangle. The term 'ragged migmatite' has been used by E. L. Boudette to describe these rocks. They also fit the description of stromatic or layered migmatites as classified by Ashworth (1985). The leucosomes are typically blebs, spots or stringers within a melanosomatic schist matrix giving it a ragged appearance. The migmatites display classic gneissic layering. The migmatites are incontrovertibly embrechitic, meaning that the gneissic layering can be tracked to indisputible bedding and that the two fabrics are parallel and essentially one and the same. This has been borne out by following

well-bedded Rangeley to migmatitic Rangeley with no change in attitude but only a change in texture and mineralogy. There are many places where calc-silicate boudins in migmatites, which are considered markers of primary layering, are parallel to gneissic layering, again demonstrating that the bedding and migmatitic foliation are similiar fabrics.

There are many possible origins for the segregations, or leucosomes, of quartz, plagioclase, plus or minus muscovite and sillimanite. They could be anatectic leucosomes, replacements after metamorphic K-feldspar, and/or augen that grew during deformation. Though most of the segregations are deformed or show at least some type of kinematic fabric that could be associated with shear zones or mylonites, it is believed that the deformation may have only enhanced the leucosomes and not produced them.

The entire Rangeley in the Gilmanton quadrangle is between 1000 and 1500 m thick. The complete thickness is not well constrained because the bottom of the formation has not been observed. The three members are all of roughly equal thickness. There are places, however, where the middle and upper Rangeley thin to only a 100 m or less in thickness. This may be due either to tectonic or stratigraphic/sedimentologic thinning. The thickness of these members varies considerably across strike; for example in the Alton quadrangle the upper Rangeley thins to only a few tens of meters wide. These thickness variations can be attributed to facies variations within the Siluro-Devonian basin.

## STRATIGRAPHIC CORRELATIONS

The stratigraphic section in the Gilmanton quadrangle is very similiar to the sequences of Silurian and Lower Devonian metasedimentary rocks described by Moench and Boudette (1970) in the CMT in west-central Maine, by Hatch and others (1983) in central New Ha mpshire CMT and by Eusden and others (1984; 1987) in the LAS. The sequence and lithologies of the sections match well. It is proposed herein that the Rangeley, Perry Mountain, Smalls Falls and Littleton Formations mapped in the Gilmanton quadrangle correlate to the formations with the same names in the LAS and KCMS of New Hampshire and the type section in the Rangeley, Maine area.

This correlation implies that the ages of the metasedimentary rocks in the Gilmanton quadrangle are Silurian and Devonian, rather than entirely Devonian, and forms a link between the units of the LAS (Eusden and others, 1984 and 1987) and the formations in central and northeastern New Hampshire along the KCMS (Hatch and others, 1983; Nielson, 1981; Thompson, 1983 and 1984; Chamberlain, 1984; G. Duke, 1984; E. Duke, 1984). In spite of the lack of paleontological and isotopic control, lithic type and sequence strongly support this correlation.

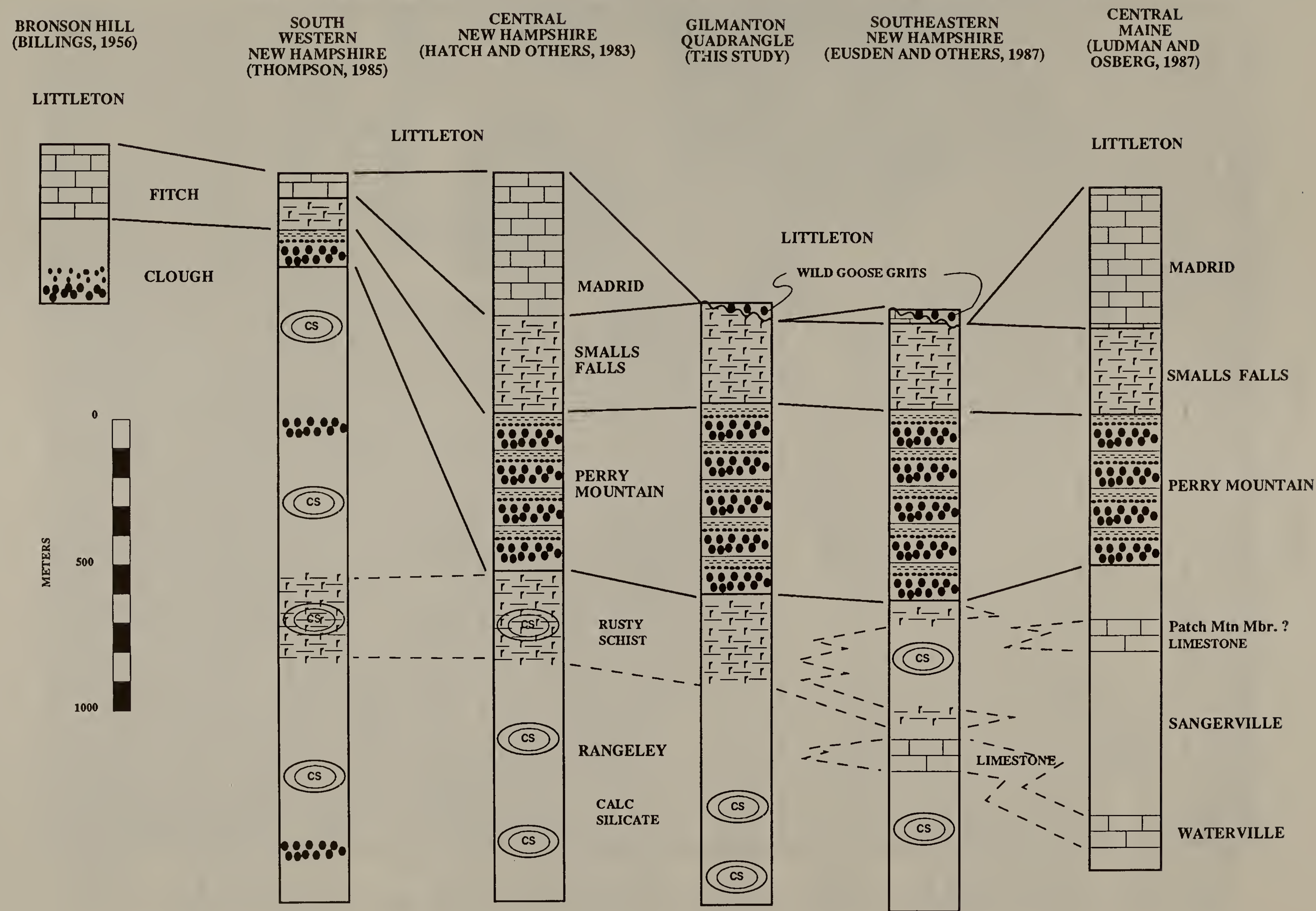
A correlation to the east with the Sangerville Formation of south-central Maine is quite likely for the middle to lower Rangeley exposed in the adjacent Alton quadrangle, where thin sequences of meta-limestones crop out. These are lithically identical to the ribbon limestone Patch Mountain member of the Sangerville. No limestones are found west of the Alton quadrangle and it would seem likely that in this area a regional facies change occurs in the Rangeley Formation where, to the east, more Sangerville-like lithologies are found and, to the west, more typical Rangeley lithologies are found. This is in accord with a recent revision of the stratigraphy by Gilman (1988) in southwestern Maine which is probably along this proposed facies transition. Figure 2 shows the stratigraphic correlations across the CMT.

## SEDIMENTOLOGY

The Silurian rocks of the western part of the CMT, near the axis of the Bronson Hill anticlinorium, are nearshore sediments (quartz pebble and polymict conglomerates and fossiliferous limestones) and to the south and east they become more distal (turbidites and euxinic shales) (Hatch and others, 1983). The transition from proximal to distal facies is in places both abrupt and gradual and has been referred to as occurring at a tectonic "hinge" (Hatch and others, 1983). In simplistic terms the hinge probably represents the transition from a shelf to a slope/rise environment.

It has also been suggested from work in Maine that the Devonian part of the section may have been derived from an easterly source (Hanson and Sauchuk, 1986; Hall and others, 1976). The boundary between the Littleton Formation and the Smalls Falls, marked by some reworking and the Wild Goose Grits, may represent the position in the section where the westerly-derived, distal, Silurian section gave way to an onlapping, easterly-derived, late Silurian-early Devonian proximal section.

Figure 2. Stratigraphic correlations across the Central Maine Terrane. Thin proximal Sillurian facies become thick distal facies to the east.



## STRUCTURAL GEOLOGY

The Gilmanton quadrangle lies in the heart of the Central Maine Terrane (CMT). It straddles structurally well-studied areas to the east (Eusden and others, 1987 and 1984) and west (Thompson, 1985; Duke, 1984; and Lyons, unpublished data). By mapping the complexly deformed metamorphic rocks in the Gilmanton quadrangle it has been possible to link structurally these areas and develop a coherent model of deformation for the CMT.

### Previous Structural Models

Billings (1956) interpreted the folds in the CMT as the product of the Devonian "Acadian Revolution." Billings' work laid the foundation for later studies and also resulted in the recognition of polydeformation, although he did not categorize the folded rocks into a regional sequence of deformations. The first effort in doing that was in the classic paper on nappes and gneiss domes along the Bronson Hill anticlinorium (BHA) by J. B. Thompson, et al. (1968), later reaffirmed by Robinson and Hall (1980) and Hall and Robinson (1982). The sequence of structural events, longstanding in the literature, starts with early, west-verging nappes which are later backfolded into east-facing structures, followed by a doming stage in which tight to isoclinal folds are produced (J. B. Thompson, et al., 1968). Adding to this scheme, P. Thompson (1985) and P. Thompson et al. (1987) argue for an early stage of thrusting that is post nappe-stage folding and pre-backfolding.

In the central portion of the CMT, Lyons and his students (Duke, 1984; Englund, 1976; Nielson, 1981; Lyons, 1979; and Lyons et al., 1982) recognized three major Acadian folding events: 1) F1, early west-verging nappes; 2) F2, broad, open folds with west or northwest-trending axes; and 3) F3, isoclinal to open folds with northeast-trending axes.

In the east part of the CMT Eusden et. al. (1984 and 1987) recognized three major folding events: 1) F1, east-verging nappes; 2) F2, tight to isoclinal folds with northeast trending axes; and 3) F3, broad, open warps with west trending axes that define a major map pattern syntaxis. Chamberlain (1985) had independently recognized the same sequence of folding in the Keene, NH area.

It is proposed here that none of these models can be uniquely applied to the entire CMT. Based on detailed mapping, not only in the Gilmanton quadrangle but elsewhere in New Hampshire by Lyons and Eusden, a new four-fold sequence of deformation is proposed: 1) F1, east and west verging nappes; 2) F2, isoclinal to recumbent folds with northeast-trending axes; 3) F3, broad, open folds with west or northwest- trending axes; and 4) F4, open to isoclinal folds with northeast-trending axes. This sequence of folding is essentially a dove-tailing of the models proposed by Eusden et. al. (1987) and Lyons et. al. (1982). Fitting into this scheme would be the Brennan Hill and Chesham Pond thrust faults of Elbert, Robinson, and Thompson (all in this volume) which developed with or after the F1 nappes and before the F2 (?) backfolds. Though these faults have not been recognized in the eastern and central portions of the CMT the arguments for their existence in the western part of the CMT are reasonable and accepted here. Other faults in central New Hampshire are not unlikely, but have not yet been identified because of the complex stratigraphy and facies changes. The field trips of Elbert, Robinson, and Thompson in this volume will review the characteristics of these faults.

### The 'Dorsal Zone'

Eusden et al. (1987) and Lyons (pers. comm., 1988) proposed that between the two major synclines in the Central Maine Terrane (CMT), the Kearsarge-Central Maine synclinorium (KCMS) and Lebanon antiformal syncline (LAS), there is an anticline, the Central New Hampshire anticlinorium (CNHA), as named by J. B. Lyons. The CNHA acts as a 'dorsal zone' splitting the CMT into two distinct structural styles. West of the 'dorsal zone' the early, F1 nappes verged west only, and east of it they verged east only; essentially mirror images of each other. The CNHA is marked not only by outcrops of the oldest formation in the CMT, the lower Rangeley Formation, but also by local 'hot spots' of granulite facies metamorphism (Chamberlain and Lyons, 1983) and an unusual trend of soapstones, presumably once ultramafic slivers injected into thinned continental crust, called the Concord Tectonic Zone (Lyons et al., 1982).

This model of the geologic structure across New Hampshire is analogous to that now recognized as "Pop Up" (Butler, 1982) or "Flower" or "Palm-Tree structures" (Ramsay and Huber, 1987) in a number of mountain belts, such as the Southern Irumide Belt, Africa (Daly, 1986). Tectonic interpretations of the Variscan Belt in Europe (from which the term 'dorsal zone' was taken) (Martin and Behr, 1983), the Caledonian Belt of Norway and East Greenland

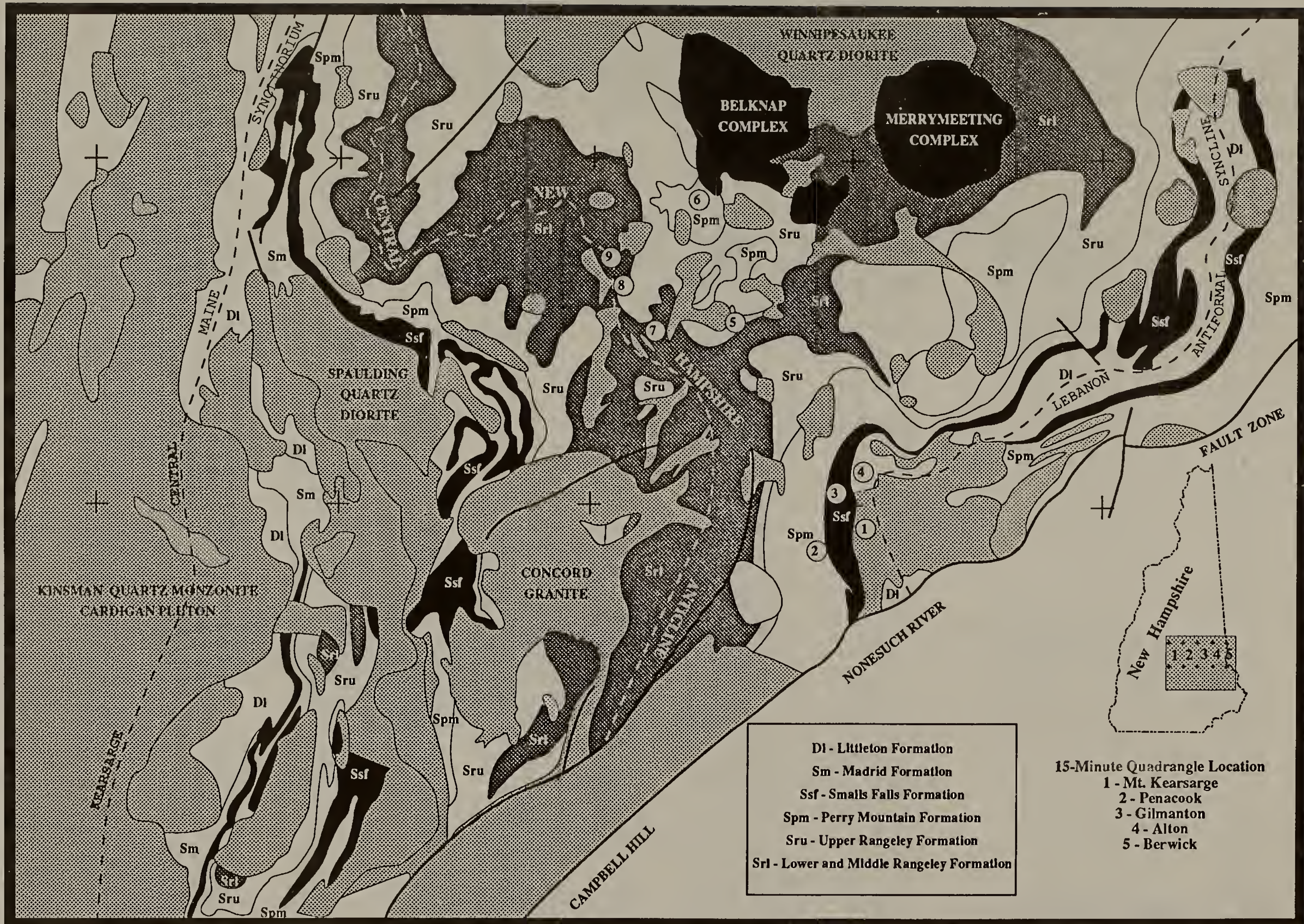



Figure 3. Geologic map of central-eastern New Hampshire. Fieldtrip stop locations shown. Modified from Lyons et. al. (1986) and Eusden (1988).

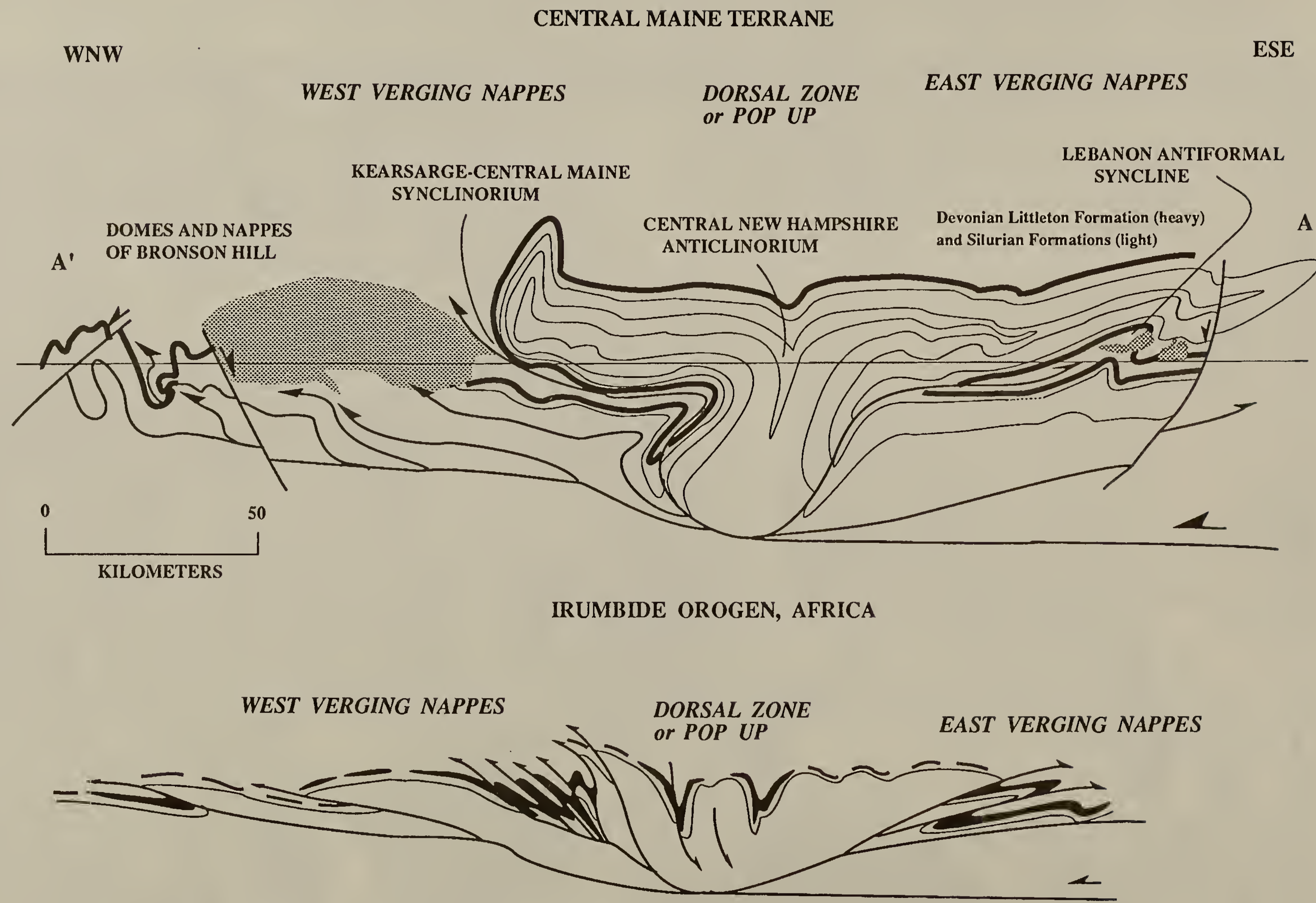



Figure 4. Cross section, located on Figure 1, through the Central Maine Terrane (modified from Lyons et. al., 1986 and Eusden, 1988) and compared to a similiar cross section through the Irumbide Orogen, Africa (after Daly, 1986).

(Ziegler, 1985) and the Grampian/Caledonian of Scotland (Thomas, 1976) also show symmetrical vergence of early thrust-nappes. All of these belts have widths roughly the same as that of the CMT, are also complexly refolded and are very similiar in cross sectional view to the model put forth here.

Figure 3 is a geologic map of part of New Hampshire detailing the area covered by this field trip and that of J. B. Lyons (this volume) through the Mt. Kearsarge and Penacook quadrangles. Figure 4 shows a schematic cross section through the region and compares it to an analogous section through the Irumbide Orogen.

The axial trace of the CNHA, or 'dorsal zone', passes through the middle of the Gilmanton quadrangle. It extends south into the Suncook and Concord quadrangles where it is truncated against the Campbell Hill-Hall Mountain-Nonesuch River fault zone. To the north it passes through the Penacook and Holderness quadrangles, eventually skirting the western edge of the White Mountain Batholith. The CNHA undoubtedly passes through the metamorphic rocks exposed in the Presidential Range but its exact location is unknown. Work is underway in that area to test the model and locate the major structures. The 'dorsal zone' with its east and west verging structures arose during the Acadian orogeny by the collaspe and tight closure of the subsiding trough of Silurian and Devonian age. The axis of this trough was not far from, if not coincident with, the axial trace of the CNHA.

## METAMORPHISM

A detailed study of the regionally metamorphosed pelitic schists had not previously been done for the Gilmanton Quadranglerangle. Adjacent areas have been well studied (Chamberlain and Lyons, 1983; G. Duke, 1984; and Eusden, 1984) and several reconnaissance studies and metamorphic maps have been made encompassing the Gilmanton Quadranglerangle (Thompson and Norton, 1968; Robinson, 1986; Lyons et. al, 1986).

Thompson and Norton (1968), Robinson (1986), and Lyons et. al, (1986) all show the Gilmanton quadrangle to be in metamorphic zone III (Tracy, 1975), the sillimanite + muscovite zone. No changes have been made concerning the metamorphic zones previously mapped in the Gilmanton quadrangle. However, intense retrograde metamorphism dominating the majority of the quadrangle has probably obscured earlier higher-grade assemblages possibly those representing zones IV, sillimanite + muscovite + K-feldspar, and V, sillimanite + K-feldspar, of Tracy (1975).

Geothermo-barometry and P–T–t path studies have been done by Eusden (1988) in this region. The following is a brief discussion of the P-T conditions during Acadian metamorphism based on this work.

Garnet core-biotite temperatures (Indares and Martignole, 1985) range from 366° to 712°C with the majority in the 500° to 600°C range. The region is characterized by hot and cold spots or plateaus with little relationship to metamorphic grade. There is a broad 600°C plateau roughly corresponding to the migmatized region, which would be expected. However, also within the migmatized zone there are several cold spots with temperatures in the 400's°C.

This type of temperature pattern was first noted by Chamberlain and Lyons (1983) and has since been found throughout the central New Hampshire CMT (G. Duke, 1984; Chamberlain, 1986; Chamberlain and Rumble, in press). Several different hypotheses have been put forth to explain the origin of the hot/cold spots.

In southwestern New Hampshire Chamberlain (1986) suggests that the distribution of synkinematic metamorphic zones (and concommitant temperatures) was controlled by folding and produced these characteristic hot and cold spots. Hot spots formed over F2 syncline-F3 syncline intersections and cold spots over F2 anticline - F3 anticline intersections.

In the Bristol, New Hampshire area Chamberlain and Rumble (in press) propose that a hot spot there, having no relation to Acadian folds, was formed by the advection of hot metamorphic fluids through the crust concentrated in narrow zones. This hotspot has an abundance of quartz veins and graphite deposits indicative of metamorphic fluid flow (see also Rumble and Chamberlain, this volume).

A third possibility is that the hot spots are spatially related to the numerous granitic sheets in the area. There is, however, no coherent relationship between the thermal spots and the map pattern of the granites.

It is likely that advection of fluids, in part constrained by major structures, was responsible for the observed temperature pattern in the study area. The garnet core - biotite 600°C hot plateau in the Gilmanton quadrangle

extends southwest through the Suncook quadrangle contiguous with the hot spot mapped by G. Duke (1984) in the Concord quadrangle. This broad hot plateau interspersed with minor hot (and cold) spots roughly follows the mapped pattern of the Lower Rangeley Formation that is exposed along the Central New Hampshire anticlinorium, the proposed root zone of the Acadian nappes. This structure would be a likely area for extensive fluid flow during metamorphism. The migmatites also show a rough correlation with this structure. This seems to reinforce the suggestion that the migmatites formed, in part, by a structurally confined flow of metamorphic fluid. However, the story must have greater complexities, as both the hot plateau and migmatite zone cut across the Central New Hampshire Anticline in a number of places.

Pressures calculated using the assemblege plagioclase-quartz-sillimanite-garnet (Koziol and Newton, 1988) have a mean of 3.6 kb. This is in excellent agreement with the observed aluminosilicates and is the best estimate of the pressure during metamorphism.

The garnet zoning profiles studied by Eusden (1988) in the Gilmanton quadrangle represent, to various degrees, the combined effects of 1) prograde growth, 2) homogenization by diffusion and 3) retrograde 'growth'. The significance of the Gibbs method P-T-t paths, calculated assuming the garnet zoning is all a product of growth zoning, is dramatically underdetermined with this in mind. At best only the P-T-t trajectories from a few of the garnets are realistic. This is supported by the observation that they generally agree with the trends defined by the geothermo-barometry and the overall four part metamorphic history outlined below.

Due to the complexities of possible thermal spikes superposed on the P-T-t paths (Chamberlain and Rumble, in press), the relationship between intersecting folds and metamorphic grade (Chamberlain, 1985), the effects of homogenization and retrograde reactions on the mineral zoning profiles, the complete P-T-t path may never be known. A single sample from the CMT may have a unique P-T-t path; certainly a handful of samples across the belt cannot adequately characterize the regions' metamorphic history. Similarly, the presumed 'peak' metamorphic temperatures and pressures calculated using garnet core compositions are subject to the same set of complications altering the original prograde garnet profile.

Metamorphic History

Based on the textures and mineral assemblages seen in the Gilmanton, Alton, and Berwick quadrangles a four-part continuum of metamorphism is proposed. The different parts are termed M1 through M4 and are segments of the Acadian P–T–t path in the region. They are not discrete metamorphic events.

M1 is a low-pressure regional metamorphism characterized by the early andalusite pseudomorphs. This is a widespread event recognized throughout the CMT.

M2 represents the highest grade of metamorphism seen in the area. The high-grade zones are probably obscurred by later retrograde metamorphism but may have reached zone IV and V, consistent with metamorphic zones mapped in adjacent quadrangles. Anatectic migmatites may have formed during M2 and metamorphic migmatites formed in association with structurally controlled fluid flux over the Central New Hampshire anticlinorium. In places, prograde garnet zoning profiles were homogenized.

M3 is a complex 'retrograde' event. Decussate muscovite, fibrolite, sympectites of muscovite + quartz and biotite + quartz, and myrmekite characterize this stage of metamorphism. During M3, garnet zoning profiles, already modified by M2, were severely retrograded. M3 assemblages represent the metamorphic field gradient seen in the region and may overprint higher-grade zones. Thermal hot spots formed by concentrated flow of metamorphic fluids and/or exsolved water from crystallizing two-mica granite sheets drove the retrograde metamorphism and may have formed metamorphic migmatites and/or altered the leucosome mineralogy of earlier anatectic migmatites.

M4 represents the waning stages of metamorphism and is characterized by scattered chlorite alteration of ferromagnesium minerals. Swarms of sericite alteration may have occurred during M4 as well.
These periods of metamorphism represent a continuum rather then a series of discrete events. They are similiar to the sequence of metamorphism proposed by Chamberlain and Lyons (1983) in central New Hampshire except their M3 has been subdivided into M3 and M4 here.

## GEOCHRONOLOGY

Monazite and sphene U-Pb ages suggest that the timing of peak high-grade metamorphism is Acadian in the CMT and Permian in the Massabesic Gneiss/Merrimack Trough. The belt of high-grade metamorphism in northeastern New England is composite, made up of crustal blocks that experienced discrete pulses of high-grade metamorphism beginning perhaps as long ago as the Precambrian and extending into the Permian. A complete discussion of these data is given elsewhere (Barreiro and Eusden, 1988 and Eusden and Barreiro, in prep.)

The metamorphic ages from the CMT in New Hampshire and Massabesic Gneiss/Merrimack Trough support the hypothesis that these are separate terranes with distinct tectono-metamorphic histories coincidentally juxtaposed at the same metamorphic grade.

The Campbell Hill-Hall Mountain-Nonesuch River fault zone, separating these two terranes, has had an active and complex history, beginning approximately ~ 360 Ma and lasting at least to 250 Ma. if not even into the Mesozoic. This boundary is a likely candidate for the Alleghanian or Variscan Front in New England.

U-Pb monazite ages from two-mica granites in the Gilmanton, Alton and Suncook quadrangles are all Devonian. This period of peraluminous granite magmatism is characteristic of the CMT. These granites are syn- to post-tectonic members of the Concord-type granitoids, part of the New Hampshire Plutonic Series. A coeval pulse of magmatism in the Massabesic Gneiss/Merrimack Trough is absent.

## SUMMARY

Figures 5 and 6 are a tectonic reconstruction of the geologic history in the Central Maine Terrane and a small portion of the Merrimack Terrane. This is an attempt to show the close association of deformation and metamorphism in this region and to pictorially synthesise the conclusions outlined above.

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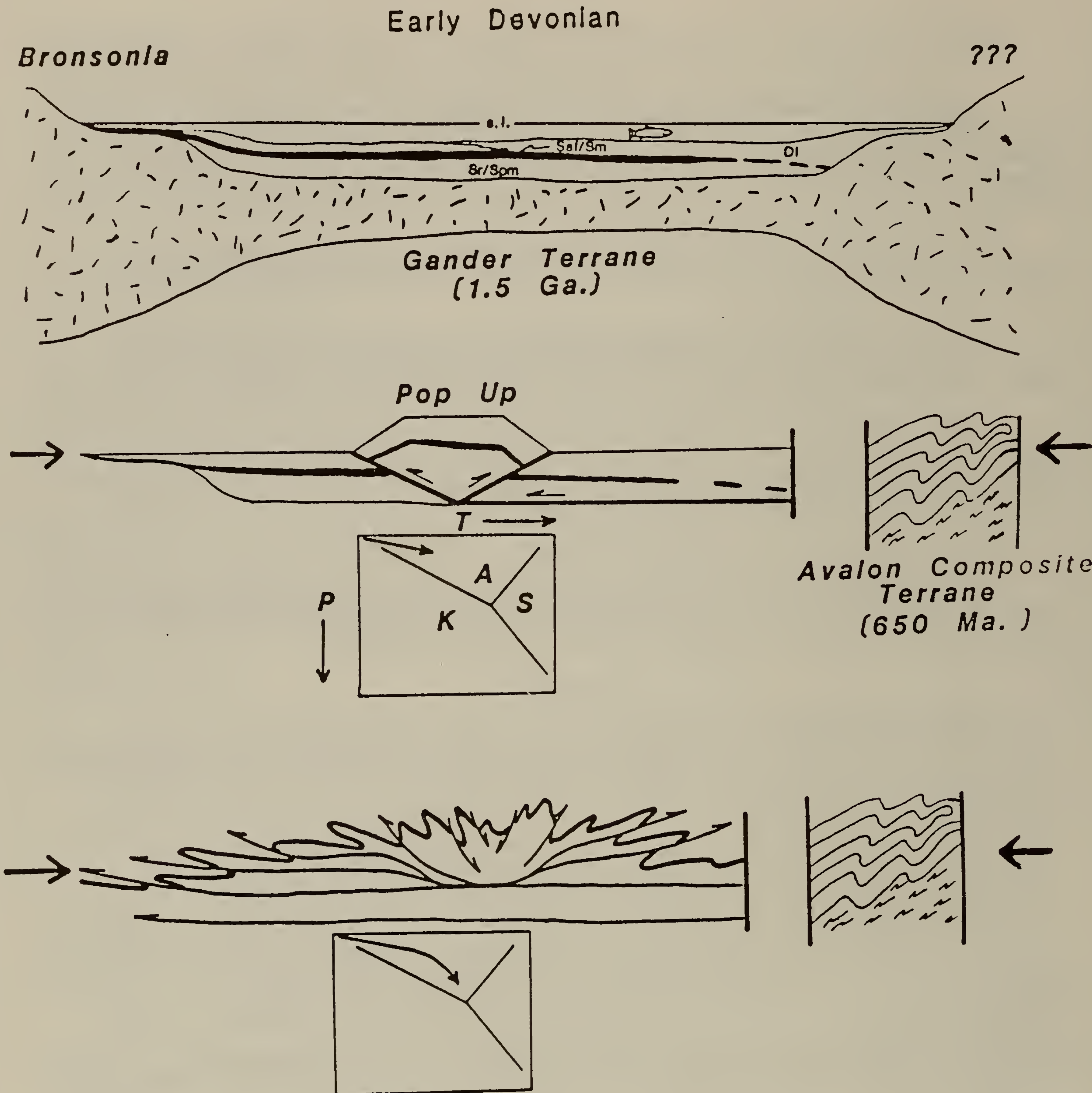


**Figure 5. Geologic reconstruction across the Central Maine Terrane in the Early Devonian. P-T path shown with respect to the aluminosilicate triple point. Compression begins immediately after deposition, Pop Up and thrust-nappes develop.**

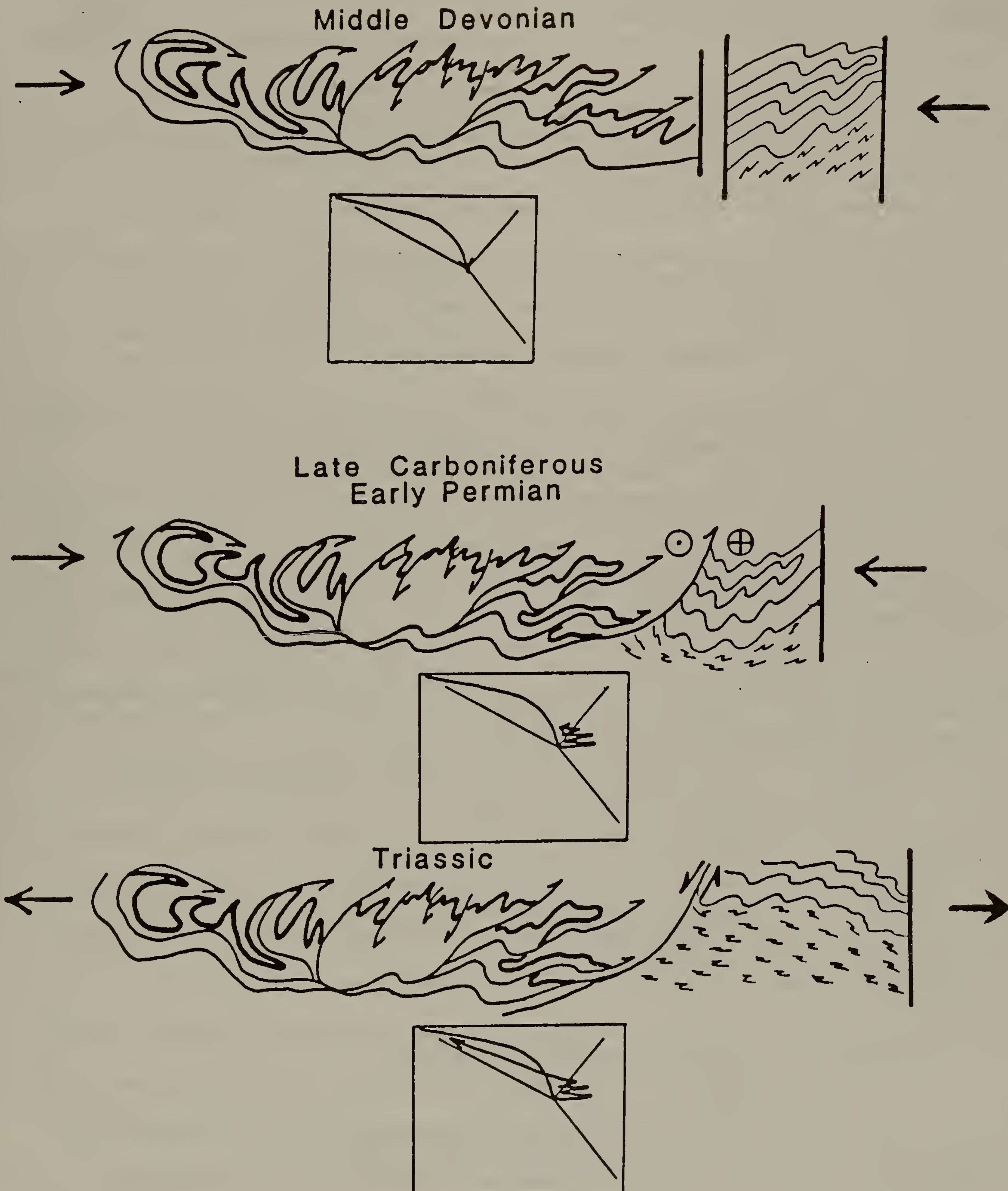


**Figure 6. Reconstruction from the Middle Devonian through the Mesozoic. Peak pressures were reached during F2. Thermal spikes occurred during and after this. Alleghanian sliver tectonics were followed by Mesozoic extension exposing basement complexes**

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## ROAD LOG

Assemble at the junction of Routes 107 and 4 in Northwood Narrows by the Minuteman RV dealer. This is the eastern of two junctions between Routes 107 and 4. Trip begins at 9:00 A.M.. Topographic maps: Gilmanton, Alton, and Suncook 15" minute quadrangles.

Mileage

0.0 STOP 1: (Suncook quadrangle) Deformed, 360 Ma. (U-Pb monazite) Barrington, two-mica granite. The S-C fabrics in this granite are associated with motion along the Campbell Hill-Hall Mountain-Nonesuch River fault zone, the terrane boundary between the Central-Maine and Massabesic Gneiss/Merrimack Trough Terranes. Was the granite deformed during Mesozoic extensional faulting or Alleghanian/Acadian thrust faulting, or both ? More work, both detailed struture and geochronology, needs to be done to unravel this faults' history.

Proceed north on Route 107 through Northwood Narrows

0.8 Go straight at top of hill turning off Route 107 which bears to the right.

1.3 Proceed straight on dirt road.

1.6 At end of dirt road where private drive turns off to the right, park on far left edge of road.

Walk along woods road, a continuation of the road we were just on, for about 10-15 minutes, passing outcrops of Smalls Falls and Perry Mountain Formation, to a beaver swamp on your right.

STOP 2: (Suncook quadrangle) Look for a nearby rocky point sticking out into the swamp. Exposed here is well-bedded upper Perry Mountain Formation caught up in a F2 downward facing antiform with well-developed minor folds and axial planar cleavage/foliation fans. These structures are on the west limb of the Lebanon antiformal syncline. Continue along woods road to clearing under the power lines. Exposed here are extensive outcrops of upper Perry Mountain. Under the lines toward the swamp one sees F4 crenulations folding either the S2 cleavage or cross beds, garnet + quartz coticules and a rare calc-silicate boudin. Immediately northwest of the road under the lines and in the woods off to the southwest are well-preserved (M1) andalumps and laminations which may be locally transposed bedding.

Return to vehicles and retrace route to Route 107, proceed north.

4.2 Bear right off of Route 107 with Jenness Pond on your right. Route 107 goes up a hill at this junction.

4.8 Take left at next next junction heading away from the pond. Road turns to dirt shortly.

5.6 At top of hill take left on dirt road marked by mailbox and park immediately on the right.

STOP 3: (Gilmanton quadrangle) Outcrops of Smalls Falls Formation. Rusty-weathering graphite and pyrrhotite bearing schist with well-developed F4 crenulations. In this area the Smalls Falls shows up as a strong positive anomaly on aeromagnetic maps. These outcrops are presumably inverted, however, there are no tops to prove it.

Turn vehicles around and return to intersection that we passed at 4.8 miles.

6.4 Take left at this junction on paved road heading northeast around Jenness Pond.

6.8 Take left on dirt road. Intersection is marked by cemetery on right.

7.5 Take left on dirt road.

8.0 Park in wide part of road at base of small hill where two private drives come into it. Views of Wild Goose Pond through trees to left.

STOP 4: (Alton quadrangle) Wild Goose Grits (lowest part of Littleton Formation). Peeled outcrops are in the woods just east of the road. An inverted section of grits with sub-rounded quartz and elongate rusty-weathering Smalls Falls clasts. Fairly good graded beds in a few localities. We are just west of the axis of

the Lebanon antiformal syncline. These grits may mark the position in the stratigraphy where the west-derived Silurian section gave way to an onlapping east-derived Devonian section.

Turn vehicles around and retrace route to the intersection where Route 107 heads west away from Jenness Pond up a hill.

10.2 We passed this junction at mileage 4.2 earlier. Proceed north on Route 107.

10.6 Smalls Falls outcrops at ponds' edge on left.

12.8 Catamount Mtn. Migmatized upper Perry Mountain and schorl pegmatite at height of land.

15.5 Take right at stop sign in Pittsfield following signs for Route 107 and Gilmanton. A good place to pick up food for lunch.

16.2 Cross Route 28. Continue on Route 107 north.

16.5 Take left to stay on Route 107.

20.8 Junction with Route 129. Continue on Route 107 north.

21.7 Park on right just beyond First Baptist Church.

**STOP 5**: (Gilmanton quadrangle) Upper Rangeley Formation. Gently dipping outcrops of rusty-weathering graphitic schist and minor quartzite. The rocks are partially migmatized here. The late large (M3) muscovite is conspicuous. The Upper Rangeley is widespread in occurrence throughout the Gilmanton quadrangle and often crops out in the low lying areas. Gene Boudette, the State Geologist, lives next door and guards these outcrops!

Continue north on Route 107.

22.5 Views of the Belknap Mountains.

24.0 Turn sharp left into the Loon Pond Beach Club private drive. LUNCH.
Continue north on Route 107.

26.3 Junction of Routes 140 and 107 in Gilmanton Corners. Continue north on Route 107.
29.4 Take right on to dirt road (Rogers Road).

30.1 Take right uphill at T intersection (Middle Route).

31.3 Park at bend in road on righthand side. Walk back about 100 yards to a tree blocked woods road on north side of Middle Route and hike northeast to the summit of Grant Hill.

**STOP 6**: (Gilmanton quadrangle) Extensive exposures of lower Perry Mountain Formation. The exposures on the southwest and highest point of the hill are chaotic, proceed northeast about 50 to 100 yards to more 'understandable' outcrops. Well-bedded calc-silicate bearing turbidite with thick quartzites typical of the Perry Mountain are exposed. Some good graded beds are seen, however, most are fast-graded and tops are equivical. F1 east-facing folds are exposed as well as F2 and F3 folds. Knots or spots of (M3) muscovite (sericite in thin section) are abundant in the pelitic layers, pseudomorphs after (M2) sillimanite. Excellent views of Mt. Kearsarge, the Belknaps, Mt. Monadnock, Mt. Cardigan, Smarts Mt., migrant raptors, and maybe some late season blueberries.

Return to vehicles and retrace route to Gilmanton Corners.

32.7 Take left back on to Rogers Road.

33.4 Take left back on to Route 107 heading south.

36.3 Take right at Gilmanton Corners (Jct. of Routes 140 and 107) heading west on Route 140 for a short bit.

36.4 Take left on Shellcamp Pond (or Meadow Pond) Road just after fire station on the left.

37.4 Shellcamp Pond on right. Outcrops of upper Rangeley and two mica granites.

37.9 Shellcamp Pond Dam. Upper Rangeley.

38.6 Proceed straight through intersection.

39.8 Take right on Loudon Ridge Road.

40.4 Park off of road on right opposite the Moore Farm.

STOP 7: (Gilmanton quadrangle) A series of outcrops in the fields southwest of the road showing extensively migmatized lower Rangeley Formation. The axial trace of the Central New Hampshire Anticline passes through these outcrops. The stromatic or layered migmatites are folded by F2 and/or F4 folds indicating that high-grade metamorphism occurred during the earliest stages of Acadian deformation. Calc-silicate and granofelsic boudins are preserved and are parallel to the migmatitic layering which is argued to be embrechitic. The leucosomes often coalesce to form early pegmatites which are in turn cut by the later two-mica granites sills and their associated pegmatites. It is possible that the migmatites formed during an intense structurally controlled fluid flux over the Central New Hampshire Anticline in the early Acadian. According to Farmer Moore the ram breeding the ewes may not like us in the same pasture and we should watch for him !

Continue northwest on Loudon Ridge Road.

41.6 Take right onto Route 106 north.

42.0 Rocky Pond on left.

42.4 Take left on South Road (dirt).

42.7 Take left on dead-end dirt road marked by "live bait" sign.

43.2 Park where road forks. There will be a sign saying "Not a through street" and another one , "Monty". Bushwack due west to the cliff about 1/2 to 1/4 mile in.

STOP 8: (Gilmanton quadrangle) Large cliff outcrop of middle Rangeley Formation. Exposed are well bedded and migmatized turbidite with no calc-silicate boudins. Bedding is dipping shallowly or, in places, is nearly flat lying. Tops here are ambiguous but the overall structure and stratigraphy suggests that the beds are upside-down. The Lower Rangeley crops out in the valley floor below us. There are several good examples of two-mica granite sills (some slightly discordant) and "bursts" of quartzo-feldspathic leucosomes. There are zones where bedding has been obliterated by localized shearing (post F1-pre F2 ?). Large late (M3) muscovite is again common and some minor F4 warps and crenulations are present. Good views from the cliff top.

Return to vehicles and retrace route to Route 106 heading north.

45.3 Heading north on Route 106. Two-mica granite sill (359 Ma. U-Pb monazite) and Middle Rangeley.

46.4 Take left for Belmont.

47.3 Take left at Getty station in downtown Belmont, back on to Route 140 heading west.

48.0 Take right into Brox Paving Materials Plant. Go past scales skirting large gravel pile on right to outcrops just north of some catch ponds, a distance of about 200 yards.

**STOP 9**: (Gilmanton quadrangle) Lower Rangeley Formation. These glacially polished coutcrops are on the axial trace of the Central New Hampshire anticlinorium, ie. the 'dorsal zone'. Bedding dips quite steeply here. Upward facing, F1 folds verge neutrally and are upright. Theoretically, just to the east the F1 folds would verge east and to the west would verge west. A good cross section of the folds can be seen on the blasted face. Follow a calc-silicate boudin-bearing horizon along strike to see the folds on the polished surface. There is a late fabric (S3/S4 ?) oblique to the beds quite visible on the blasted face.

**End of trip.** To get to Keene, proceed west on Route 140 to Tilton then take I-93 south to Concord. Take I–89 north from Concord to Hopkinton then take Route 9 all the way to Keene. Driving time about 1 hour.

# GEOLOGY OF THE PENACOOK AND MOUNT KEARSARGE QUADRANGLES, NEW HAMPSHIRE

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## REGIONAL GEOLOGIC SETTING

In order to clarify what is now a rather confusing set of tectonic labels in northern New England, Zen (1983, and ms.) has proposed that the geologic terrane lying between the Monroe fault on the west and the Campbell Hill fault on the east be termed the Central MaineTerrane. This suggestion has been followed both in this article, and in the accompanying article by Eusden (this volume) on the Gilmanton quadrangle. The reasons are that a terrane is identifiable chiefly on the basis of the age of its basement, rather than that of its cover, and that it should have well demarked boundaries.

In the Central Maine Terrane the basement, exposed only in the 1.5 Ma Chain Lakes allochthon of northwestern Maine, is different petrographically and older geochronologically than the Grenvillian rocks west of the Connecticut River (Aleinikoff and Moench, 1985; Lyons, Aleinikoff and Zartman,1986; Harrison, Aleinikoff and Compston, 1987). There are also scattered serpentinite pods along the Connecticut Valley, tracing northerly toward the Boil Mountain ophiolite and the extensive Hurricane Mountain melange in Maine. Just to the east is what is now generally regarded as the leading edge of a paleo-volcanic arc, the Bronson Hill anticlinorium. As proposed by Lyons, Boudette, and Aleinikoff (1982), and Zen (1983), the Connecticut Valley region is at least as likely to have been the site of the closure of Iapetus as the Baie Verte-Brompton Line, which lies entirely within Grenvillian basement.

Toward the southeast, structures of the cover rocks of the Central Maine Terrane are decapitated along the Campbell Hill fault by the pre-Early Ordovician Massabesic-Merrimack-Rye Terrane of Eusden, Bothner, and Hussey (1987), in which there have been several reports of 650 Ma ages (Besancon, Gaudette and Naylor, 1977;Aleinikoff, Zartman and Lyons, 1979; Kelly, Olszewski and Gaudette, 1980).

Within the Central Maine Terrane, from west to east, are the Bronson Hill anticlinorium, the Kearsarge-Central Maine synclinorium, the Central New Hampshire anticlinorium, and the Lebanon antiformal synclinorium. This excursion commences in the Central New Hampshire anticlinorium and progresses southwesterly into the Kearsarge-Central Maine synclinorium. The concurrrent trip to the east, by Eusden ( this volume) will mirror-image the stratigraphy and structure because vergence is westerly west of the anticlinorium and easterly east of the anticlinorium (Eusden, Bothner and Hussey, 1987).

Eusden (1988) and my current hypothesis concerning the major structural pattern relates, in part, to the Silurian-Early Devonian paleo-geology. During that time this region was a deeply-subsiding trough receiving an accumulation of more than 4 km. of westerly-derived Silurian clastics and volcanics, and at least 1 km. of easterly-derived Early Devonian flysch ( Hatch, Moench and Lyons,1983). The basement was evidently extending during this depositional cycle, but toward the end of the Siegenian Epoch underwent compression (the Acadian orogeny), and the trough was inverted into a "pop-up" structure (Butler,1982) analogous to those in the Variscan range of western Europe, and the Caledonides of East Greenland and Scandinavia (see also Eusden, this guidebook).

An interesting, but occult, question is why there are such voluminous injections of sheet-like Acadian plutons (408 to 365 Ma) in this terrane. Their tabular forms (Nielson and others, 1976) are understandable if some of them are synkinematic (Kinsman and Bethlehem suites) or late-kinematic (Spaulding suite), and are involved in nappe or thrust-nappe tectonics, as they are. But their timing is an enigma. Thermal calculations by Chamberlain and England (1985) show that tectonic thickening followed by recovery of the isotherms could produce conditions in the lower crust suitable for the generation of anatectic melts, but only 50 to 100 m.y. subsequent to tectonism. This scenario might possibly apply to the 365 Ma Concord suite, the 325 Ma Sebago and Effingham intrusives of eastern New Hampshire, or possibly the 275 Ma granite at Milford in southern New Hampshire, but surely not to the more abundant earlier intrusives. One is left with the intuitive thought that extension tectonics and crustal thinning may solve the problem, but among the thermal experts who have considered the matter, this theory likewise fails (C. P. Chamberlain, verbal communication), at least for the crustal lithosphere. Nevertheless, mantle-type initial Sr

isotopic ratios in Kinsman and Spaulding suites (Lyons and Livingston, 1977), imply that the upper mantle and/or asthenosphere are implicated in the generation of magmas and the upward advection of heat, possibly even concurrently with the Silurian extensional event. Extension in the mantle may have initiated pressure-release melting. The ascent and ponding of those melts in the lower crust may, in turn, have created the conditions necessary for the generation of the anatectic granitoids which constitute the New Hampshire Plutonic Series. Injection of sheets of these intrusions into the middle crust may then have created the requisite conditions for regional metamorphism.

## METAMORPHISM

The same thermal problem exists in explaining the early Acadian advent of regional metamorphism, which has been dated in these environs by Barreiro and Eusden (1988) at 403-386 Ma, using Pb/U isotopic systematics on monazites from the metapelites. These ages overlap, quite closely, the isotopic ages determined on the Kinsman, Bethlehem, and Spaulding Intrusive Suites, suggesting a causal relationship between magmatism and regional metamorphism. That explanation has been advanced, for example, for metamorphism peripheral to the 325 Ma Sebago batholith of Maine and eastern New Hampshire (DeYoreo, Lux and Guidotti, in press). On the other hand, much of central New Hampshire lies within the sillimanite-muscovite zone of regional metamorphism, with rather abrupt gradations upward through sillimanite-muscovite-Kfs, and sillimanite-Kfs zones into granulite-facies sillimanite-garnet-Kfs-cordierite "hot spots" (cf., fig. 3). However, from the mapped geology and what has been inferred about the near-subsurface by gravity measurements, there is no close correlation of the "hot spots" with intusives in central New Hampshire, so the cause of the overall metamorphic fabric cannot be said to be understood. Chamberlain (1987) has explained the hot spot pattern in southwestern New Hampshire in terms of repeated folding of isotherms during regional metamorphism, but Chamberlain and Rumble (1986) also look favorably on the advection of hot fluids as a possible cause for the "hot spots". The problem remains open.

In the Penacook and Mt. Kearsarge quadrangles the metamorphic pattern (fig. 3) is complex. Although the muscovite-sillimanite zone dominates, there are abrupt and irregular shifts to pockets of lower-grade,staurolite-muscovite rocks, and to higher-grade sillimanite-Kfs and granulite facies assemblages, in unpredictable patterns. The $Al_2SiO_5$ isobar crosses bothquadrangles in a northeasterly direction, with the higher-pressure (> 3.75kb ) rocks to the northwest. The Kinsman Quartz Monzonite an early synkinematic intrusive underlying much of the Mt. Kearsarge quadrangle was subjected to regional metamorphism, but has been shown by Plank (1987) to have a relict mineralogy which even preserves, in its garnet-biotite pairs, relict temperatures in the 750-900° C range. These contrast with the 500-700°C temperatures in the wallrocks.

Chamberlain and Lyons (1983) identify at least three stages of Acadian metamorphism : 1) M1 andalusite metamorphism evidenced by sillimanite pseudomorphs after andalusite, and by flecky gneisses at contacts of intrusives such as the Cardigan pluton of Kinsman Quartz Monzonite; 2) M2 overprinting of M1, generally upgrade, and responsible for the isograd patterns on fig. 3, and 3); M3 local retrogradation. In at least some field situations, the retrogradation is ascribable to fluid circulation peripheral to Concord Granite plutons.

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FIG. 1

GEOLOGIC MAP OF THE MT. KEARSARGE QUADRANGLE

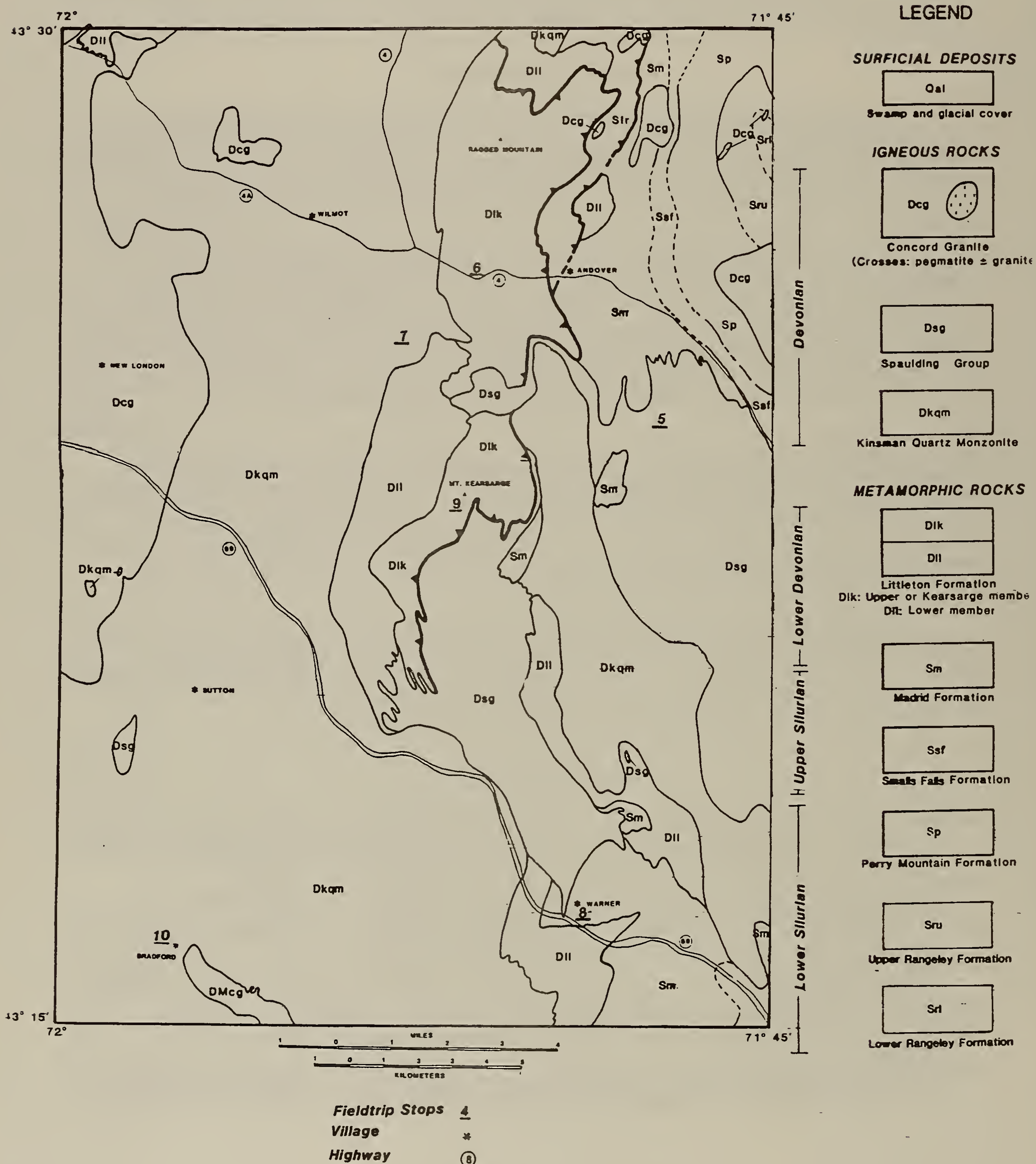
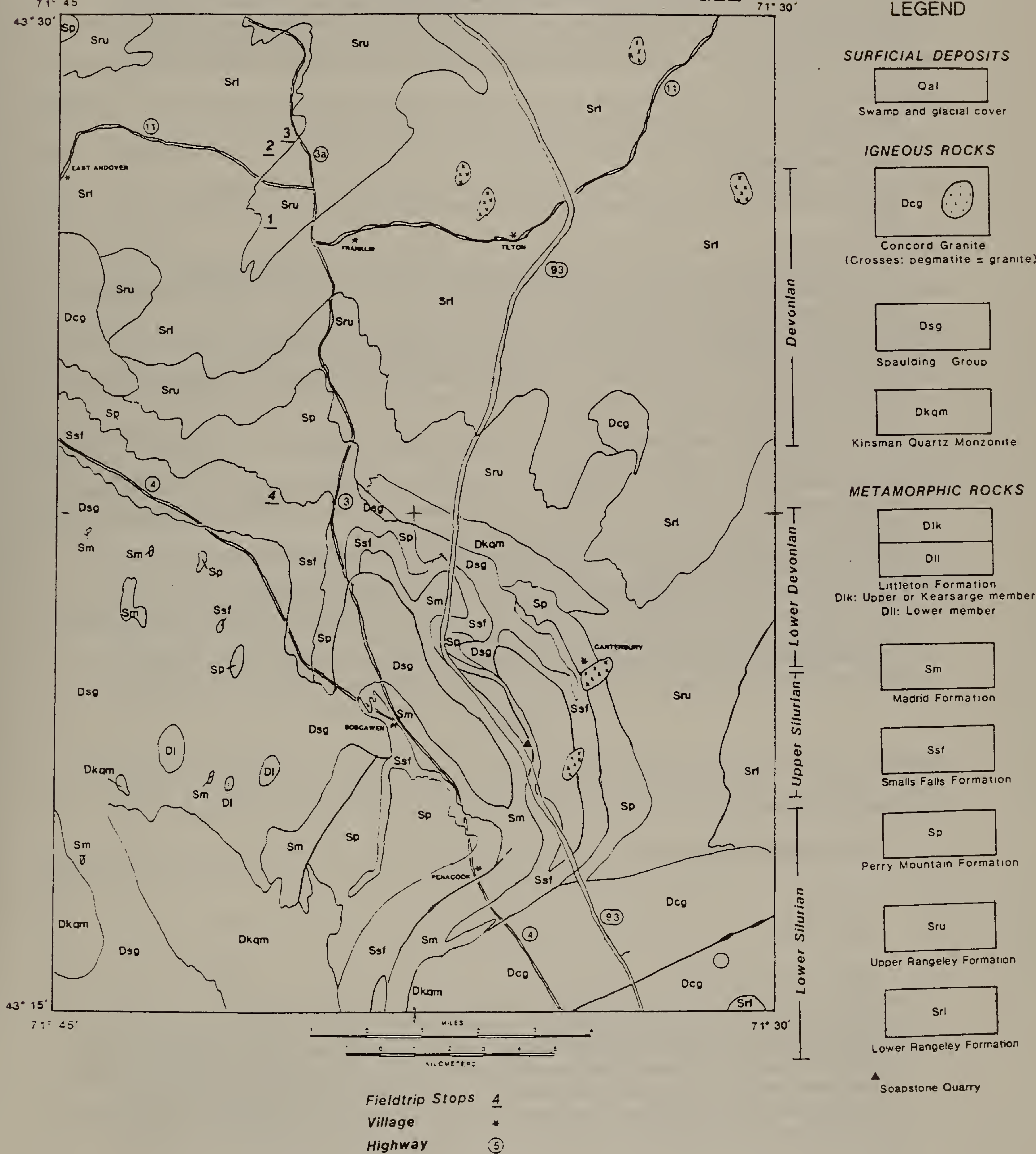


FIG. 2

## GEOLOGIC MAP OF THE PENACOOK QUADRANGLE



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## ITINERARY

ASSEMBLY POINT: Junction of Rte 3A and Rte 11 north of Franklin, on the west side of the Pemigewasset River.
Proceed westerly on Rte 11.

0.4 Left just beyond cemetary

0.8 Left again

1.3 Right up hill

1.8 STOP 1 , on powerline

Nearer the road are outcrops of rusty-weathering gar-sill-bio-musc schists with some calc-silicate boudins, typical of the upper Rangeley Formation. Further along the schists become "lumpy", and then less rusty, well-bedded, and coticule-bearing. Numerous leucosomes, whether from anatectic melting or metamorphic segregation, become conspicuous. We have crossed into lower Rangeley Formation through a typical transition zone. In dealing with structural features of the bedrock, the long-established convention of calling

bedding S0, first foliation S1, second-generation foliation S2, folds of the first generation F1 etc.will be followed. Structures in these outcrops are chiefly S1, striking approximately N 60 E. There is also superposed folding, F3, deforming S1, and plunging easterly. Looking toward the northeast, across the Pemigewasset Valley are Sanbornton and Hersey Mountains in the Holderness quadrangle. These are held up by a lentil of well-bedded and graded-bedded lower Rangeley deformed by well-developed westerly-plunging F3 folds. The trace of the Central New Hampshire anticlinorium coincides with the outcrop of the lower Rangeley. To the east, F1 structures verge easterly ; to the west, westerly.

Turn around and return to Rte 11.

3.2 Left turn on Rte 11

3.3 Right on Lake Ave.

4.0 Right at T

4.5 STOP 2 . Park on left side of road and walk back to outcrop.

Sharply, but thinly interbedded quartzite and metapelite typical ofmuch of the lower Rangeley. It is beds such as these, but thicker,which underlie Sanbornton and Hersey Mountains to the northeast.Bedding strikes N 45 E, dips 64 SE, and is upward-facing toward the projected trace of the upper-lower Rangeley contact to the southeast.

Return to cars

4.6 Right turn

4.9 STOP 3 on powerline

"Lumpy" coticule-bearing outcrops, resembling some of those atStop1, but here graded bedding is also evident. Rocks such as those at Stop 2 are exposed to the northwest along the powerline.Grades nearest the road face southeasterly toward mapped upperRangeley, but a little further along they face northwesterly, indicating, to no one's surprise, that the formation here, as elsewhere,is isoclinally folded and has S0 and S1 fabrics. Almost everywhere in this region the earlier fabrics are deformed by westerly to southwesterly-plunging F3 folds.The origin of the lumps is not totally understood. Mineralogicallythey consist of concentrations of gar+bio+sill+stau+plag+qtz, and may be pseudomorphs of staurolite porphyroblasts which have been destroyed by an upgrade stau+musc reaction.

5.1 Right turn on Rte 3A

5.3 Outcrops in road cut are somewhat rusty metapelite with abundant calc-silicate boudins, and are mapped as upper Rangeley.They would lie on the southeastern flank of the Central New Hampshire anticlinorium.

Proceed southerly past Franklin, where the Merrimack River begins at the confluence of the Pemigewasset and Winnipesaukee Rivers.

7.9 Right on Rte 127

12.6 Left on dirt road

14.3 STOP 4 Proceed northerly uphill for approximately 600 feet through a completely inverted sequence of Small Falls at the base of the hill, and Perry Mountain toward the top. Smalls Falls is extremely rusty-weathering, and consists of essentially two lithologies, one of these a fine-grained highly pyrrhotitic calc-silicate, the other a porous white qtz-plag-musc-phlog rock, with conspicuous graphite. PerryMountain consists of sharply interbedded quartzite and metapelitein which ( unlike the Littleton Formation) it is difficult to decipherfacing directions.Formation boundaries striking approximately N 50 W trace from the Penacook into the Mount Kearsarge quadrangle over a distance of seven miles, and here verge strongly toward the southwest. At either end of this belt they resume a northeasterly trend, and westerly dips.

Throughout much of western New Hampshire, on both map and outcrop scale, there is late-stage sinistral deformation, of which this belt is apparently one manifestation. In eastern New Hampshire,however, the map pattern implies late-stage dextral deformation.We relate these deformations to the F3 or F4 folds.One of the Perry Mountain outcrops shows a minor F1 fold, trending N 50 W, with a sense-of-shear implying a downward facing ( i.e., toward the Smalls Falls ). Much more obvious is the fact that the S0-S1 structures, which were originally sub-horizontal, have been deformed by open northeasterly-trending F4 folds.

Turn cars around, and return to Rte 127

16.0 Left on Rte 127

17.0 Right (west) on Rte 4. Poorly exposed outcrops along Rte 4 for a few miles are Smalls Falls.

19.0 Left on West Salisbury Road, just past settlement of Salisbury Heights . DANGEROUS TURN.

21.0 Left at foot of hill

21.4 **STOP 5** Park at bridge across Blackwater River.

Outcrops in the bed of the river are in the contact zone of the areally extensive, but thin (< 1 km) Blackwater pluton. The predominant rock type is a spotted weakly to moderately foliated biotite granodiorite or tonalite, with gradations to granite (Duke,1978). Garnet is present in many of the biotite clots, and two-mica granite is common, but some rocks contain hypersthene or pargasite,so the pluton is both peraluminous and metaluminous. To the south,in Webster, this pluton crosscuts the Cardigan pluton, engulfing large xenoliths of Kinsman Quartz Monzonite. It is considered to be a part of the Spaulding Intrusive Suite ($\pm$ 392 Ma). The numerous xenoliths are chiefly calc-silicates which arerefractory, whereas their original enclosing metapelitic host has been assimilated by the pluton. The host formation is an unusually rusty and calc-silicate rich facies ("Andover Member") of the Madrid Formation.

Retrace route

21.8 Left at fork in road

23.6 Left onto Rte 4. DANGEROUS INTERSECTION

25.5 Junction of Rtes 4 and 11. Outcrops here strongly resemble upper Rangeley, but Madrid outcrops occur to the west and Smalls Falls to the east. If the outcrops are Rangeley, there would have to be a folded thrust, with this a fenster exposing the lower plate. This is not an impossibility, but there is no field evidence to support this interpretation.

28.4 **STOP 6** Well bedded and well-graded metaturbidite of the Littleton Formation, characteristic ( but not absolutely diagnostic) of its upper member. A series of fairly tight upward-facing F4 folds can be deduced by walking across the top of the outcrop. Quartz vein at eastern end of outcrop is liberally laced with tourmaline, the boron for which may have been freed from the pelites themselves during progressive regional metamophism.

29.1 Junction of Rte 11 and Rte 4A . Stay on 11

30.2 **STOP 7** Garnetiferous phase of Kinsman Quartz Monzonite ( 408 $\pm$ 5 Ma) toward the eastern edge of Cardigan pluton. As is typical, there are large megacrysts of partly sericitized K-feldspar mantled by myrmekite, the garnet is rimmed or replaced by biotite, xenoliths foliation. Clark and Lyons ( 1986 ) have shown that Kinsman compositions range from tonalite to granite, that the K-feldspars cannot have crystallized from an initially homogeneous melt, that the garnet is igneous, that the plagioclase also has igneous characteristics, and that geochemically Kinsman compositions can be accounted for on the basis of varying mixtures of leucogranite, assimilated pelite (as displayed, for example,at Stop 5), and restite. Plank (1987) found that the 750-900$^0$ C garnet-biotite temperatures in the Cardigan Pluton are higher toward the margins, just as Clark and Lyons (1986) report that the K-feldspars within the pluton are annealed to maximum microcline, whereas those in the wallrocks are orthoclase. The xenoliths here are probably of the

FIGURE 3. GEOLOGY OF SUMMIT AREA, MT. KEARSARGE

Madrid Formation, buttressing the idea that Mount Kearsarge is essentially a folded recumbent syncline, opening westerly.

Proceed on Rte 11

31.2 Left off Rte 11

35.8 Right at T intersection

36.3 Left toward I-89

36.4 Left onto I-89 South

36.9 Public Rest Stop

43.7 Offramp from I-89. At end of ramp turn left to Warner on Rte 103

44.6 Right turn on Mill Street

44..8 STOP 8 Park cars and walk to outcrops on south side of I-89.

Bio-plag-qtz granofels with zoned calc-silicate boudins in the Madrid Formation (originally, but unofficially, dubbed the Warner Formation). Note the distinctly rustier outcrops to the east, several apophyses from a granodioritic (Spaulding ?) pluton which underlies much of the area between Warner and the summit of Mt. Kearsarge, and the spheroidally-weathered, columnar-jointed diabase dikes. At least two generations of folds are present. Earlier F1 folds can be seen in these outcrops, but are best displayed on the flat I-89 median strip, which we cannot visit. Here the F1 folds have a bearing of $240^{o}$, and plunge 53 SW, in approximate conformity with steeply plunging folds on the north side of this outcrop, and with lineations visible on the underpass cuts. Later open folds F4(?) plunge gently toward N 20 W.

Turn cars around, and return up Mill Street.

45.0 Right turn

45.1 Left turn toward Rollins State Park

49.3 and Kame on right at approximately 900-feet elevation. More extensive gravelly kames occur on both the east west slopes of Mount Kearsarge at elevations of up to 1200 feet.

50.1 Entrance to Rollins State Park. Narrow climbing curvy road.

53.6 STOP 9 Parking lot. Proceed up trail toward summit.

The summit area affords a superb display of graded-bedded Littleton Formation along and on either side of the trail, and unequivocal examples of several reversals in facing directions. This is the axial zone of the Kearsarge-Central Maine synclinorium. The Littleton Formation here is lithologically indistinguishable from the Seboomook Formation of western Maine Bedding and foliations dip steeply outward on both the eastern and western flanks of the mountain, and flatten toward the summit, giving the impression is of a northerly-plunging antiform. Mapping on a 1:50 foot scale reveals a more complex pattern. A series of east-northeasterly trending axial planes of early recumbent folds (S0 = S1) have fold hinges (F1 or F2) plunging northeasterly. Deforming these structures are a series of northwesterly-plunging (F 3?) and northeasterly-plunging ( F4 ) folds. Many of the folds are downward-facing. The major structure on Mt Kearsarge is related areally to the geology at Stop 4, and at the summit is apparently chiefly on the inverted limb of a westerly-verging refolded F1 nappe. The Cardigan and Weare plutons very nearly join north of Mount Kearsarge. They were probably originally part of the same plutonic sheet which has now been folded and eroded.

Return to Warner

62.3 Right turn in Warner onto Rte 103 toward Bradford.

63.7 Just beyond the I-89 underpass is an elegant outcrop of Kinsman displaying a folded aplite with axial planar cleavage, as well as garnet replacement by biotite as one approaches the aplite.

70.2 Intersection with Rte 114. Outcrop on right just before the stoplight is Lake Massasecum tonalite, a part of the post-tectonic Concord Intrusive Suite. On the lake shore, the tonalite has excellent flow banding.

71.3 <u>**STOP 10**</u> Roadcut west of the stoplight in Bradford village.

Garnetite pod and layer within the Kinsman. Relations on both sides of road show that the garnetite ( gar-cord-bio-plag-qtz) isa restite from a slab of Littleton which is undergoing anatexis. The anatectic melt is diffusing into the pluton. Gravity studies indicate that the Cardigan pluton here is at its maximum preserved thickness of 2.5 km. .

END OF FIELD TRIP

Best route to Keene from here is to return through Bradford to Rte 114. Turn right (southeast) on Rte 114 , which, after 9 miles, intersects Rte 9. Turn right (southwest) on Rte 9, and stay on it to Keene.

# TECTONIC AND METAMORPHIC EVOLUTION OF THE BERNARDSTON NAPPE AND THE BRENNAN HILL THRUST IN THE BERNARDSTON-CHESTERFIELD REGION OF THE BRONSON HILL ANTICLINORIUM

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## INTRODUCTION

In the central New England Appalachians crystalline bedrock is exposed in three broad structural features. These belts, from west to east, are: the Connecticut Valley-Gaspé synclinorium, the Bronson Hill anticlinorium, and the Merrimack synclinorium (Hall and Robinson, 1982; Robinson and Hall, 1980; Zen and others, 1983). Central to this field trip is the Bronson Hill anticlinorium, a region characterized by a northward-trending series of mantled gneiss domes (Thompson and others, 1968) in which pre-Silurian rocks are exposed.

In the Bronson Hill anticlinorium, the oldest rocks are late Precambrian metamorphosed rhyolites and interbedded sedimentary rocks exposed in the core of the Pelham dome (Robinson, 1963; Zen et al., 1983). Late Precambrian(?) to Middle Ordovician(?) rocks occur in the outer part of the Pelham dome and in the cores of the other gneiss domes within the Bronson Hill anticlinorium. Unconformably(?) overlying the gneisses in the domes are the Middle Ordovician(?) Ammonoosuc Volcanics and the Partridge Formation. These two lithostratigraphic units are the remains of a volcanic arc complex. The Ammonoosuc contains metamorphosed tholeiitic basalts and andesites, dacites, and rhyolites, interpreted to have been produced from melting of a mantle or basaltic source (Aleinikoff, 1977; Schumacher, 1981, 1983, 1988; Leo, 1985). The Partridge Formation is comprised of graphitic, sulfidic schists interbedded with metamorphosed mafic and felsic volcanics geochemically similar to arc volcanics (Hollocher, 1985).

Along the Bronson Hill anticlinorium these rocks are unconformably overlain by a sequence of fossiliferous Silurian-Devonian units that were described by Billings (1937) in the Littleton, N.H. area. The sequence contains three distinctive units. Stratigraphically lowest of these is the Silurian Clough Quartzite. The Clough is largely a metamorphosed conglomerate with deformed vein-quartz pebbles in a quartzite matrix and is discontinuously overlain by the marbles and calc-silicate granulites of the Silurian-Devonian Fitch Formation. The Lower Devonian Littleton Formation contains interbedded quartzites and gray weathering, pelitic schists. The Littleton is the youngest unit in the field-trip area to have been deformed and metamorphosed during the Acadian orogeny.

The relatively thin Silurian sequence found in the Connecticut Valley belt is interpreted to be the proximal equivalent of the much thicker sequence of stratified rocks found just to the east in the Merrimack synclinorium (Hall and Robinson, 1982) which has been correlated with units in western Maine (Hatch and others, 1983; Thompson, 1985, this volume) . This eastern stratigraphic sequence is made up of the Rangeley, Perry Mountain, Francestown, and Warner formations. The distinctive succession of varied rock types within the eastern sequence facilitates correlation between field areas and is the key to recent structural interpretations.

Paleozoic rocks in the region are unconformably overlain by a sequence of Triassic-Jurassic sedimentary and volcanic rocks (Zen and others, 1983) deposited in the basins associated with the rifting of North America from the Pangaea supercontinent. The sedimentary rocks comprise the Newark Supergroup and include fanglomerates deposited in alluvial fans against the scarps of normal faults on the sides of the

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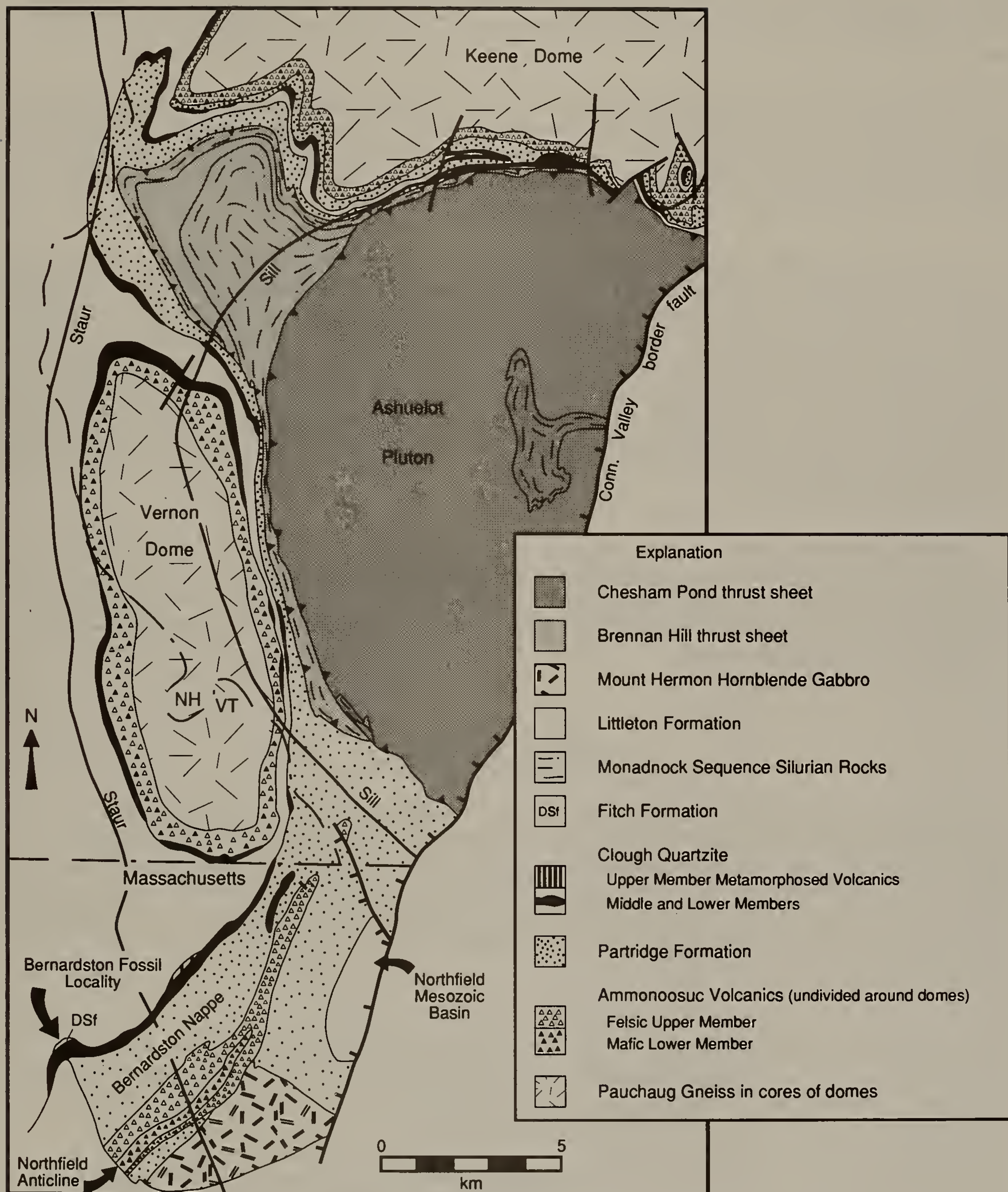



Figure 1. Geologic map of the Bernardston-Chesterfield area compiled from Trask (1964), Elbert (1984, 1986, 1987), and Hepburn and others (1984). Tectonic levels of Brennan Hill and Chesham Pond thrust sheets shaded in gray; rocks in Ashuelot pluton, Mesozoic basins, and east of Connecticut Valley border fault not patterned. Staurolite- and sillimanite-in isograds shown by solid lines with abbreviation for index mineral on high-grade side. Mesozoic faults are shown by hatchured lines; hatchures on downthrown side.

basins, fluvial deposits, and lake and swamp deposits. Basaltic flows and sills are present in the basins, while associated dikes intruded the nearby crystalline rocks. Igneous activity continued in the Cretaceous with the formation of the alkalic ring dikes and lavas of the White Mountain Magma Series, as well as more widespread, mafic dikes.

The Bernardston-Chesterfield area is along the western edge of the Bronson Hill anticlinorium, at the northern limit of the Mesozoic Deerfield basin on the down-thrown, western block of the Connecticut Valley border fault (Figure 1 of Robinson, trip C-4, this volume). The area lies on the southern, eastern, and northern borders of the Vernon dome, one of the series of en-echelon Acadian (Devonian) gneiss domes of the Bronson Hill anticlinorium (Thompson and others, 1968).

Acadian deformations in central Massachusetts and adjacent New Hampshire have been summarized by Thompson and others (1968), Robinson (1979), Robinson and Hall (1979), and Hall and Robinson (1982). These are broadly divided into an early nappe stage, an intermediate backfold stage, and a late gneiss dome stage. The nappe stage involved the formation of west-directed fold nappes. These nappes were folded back towards the east during the backfold stage and then folded again during the rise of the gneiss domes in the Bronson Hill belt. The recent correlation and mapping of the distinctive, eastern Silurian-Devonian sequence and its relation to the thinner, western sequence, has led to the recognition of regional-scale thrust faults that cut the fold nappes of the Bronson Hill anticlinorium and are then folded by the regional backfold and dome-stage phases of Acadian deformation (Thompson, 1985; Elbert, 1986; Robinson, 1987; Thompson and others, 1987).

In the Bernardston-Chesterfield area (figure 1) one such thrust fault has juxtaposed a section of the eastern, Monadnock-western Maine sequence against the Partridge Formation in the core of the Bernardston nappe (Elbert, 1986; Thompson and others, 1987). This thrust fault is correlated with the Brennan Hill thrust, proposed on somewhat less compelling evidence grounds in the Monadnock quadrangle by Peter Thompson (1985 and this volume).

## TRIP OBJECTIVES

This trip is designed to present the stratigraphic and structural evidence for the Brennan Hill thrust while highlighting several important aspects of the early tectonic development of the Bronson Hill anticlinorium. The route traverses from rocks deposited west of the "tectonic hinge" (Hatch and others, 1983), in the Bernardston nappe, to rocks that were deposited east of the "tectonic hinge", at Biscuit Hill, Hinsdale, N. H. Detailed stratigraphic mapping is the most powerful tool available to help understand the early structural geometry and tectonic history of an orogen and the itinerary is built to highlight the stratigraphic relationships in the area.

**Geology of the Bernardston Northfield Area** (figure 2): A geologic map of the southern part of the field-trip area is presented in figure 2. The rocks of the Bernardston area have been previously interpreted as lying in a homoclinal sequence on the inverted limb of a regional fold nappe (Trask, ms; Trask and Thompson, 1967; Thompson and others, 1968; Zen and others, 1983). Thompson and others (1968) interpreted the nappe at Bernardston as a continuation of the Skitchewaug nappe to the north. Subsequently, Thompson and Rosenfeld (1979) suggested that rocks previously considered part of the Skitchewaug nappe were, instead, parts of two separate nappes: an upper nappe named for Skitchewaug Mountain, Vermont, and a lower nappe, the Bernardston nappe, continuous with the section near Bernardston, Massachusetts.

My recent work in the Bernardston-Northfield area (Elbert, 1984) has produced stratigraphic and structural evidence supporting substantially new interpretations and regional correlations. The map in figure 2 shows that much of the eastern part of the area is underlain by Ordovician strata which area exposed in the core of the West Northfield isoclinal anticline. Rocks now assigned to the Ammonoosuc were previously mapped as volcanics in the Partridge Formation (Trask, ms; Trask and Thompson, 1967; Thompson and others, 1968; Zen and others, 1983). These rocks were reassigned based on stratigraphic correlation of three mapped members within the Ammonoosuc in Bernardston with other Ammonoosuc sections around the southern part of the Keene dome (fig. 1 of Robinson, trip C-4, this volume) and the

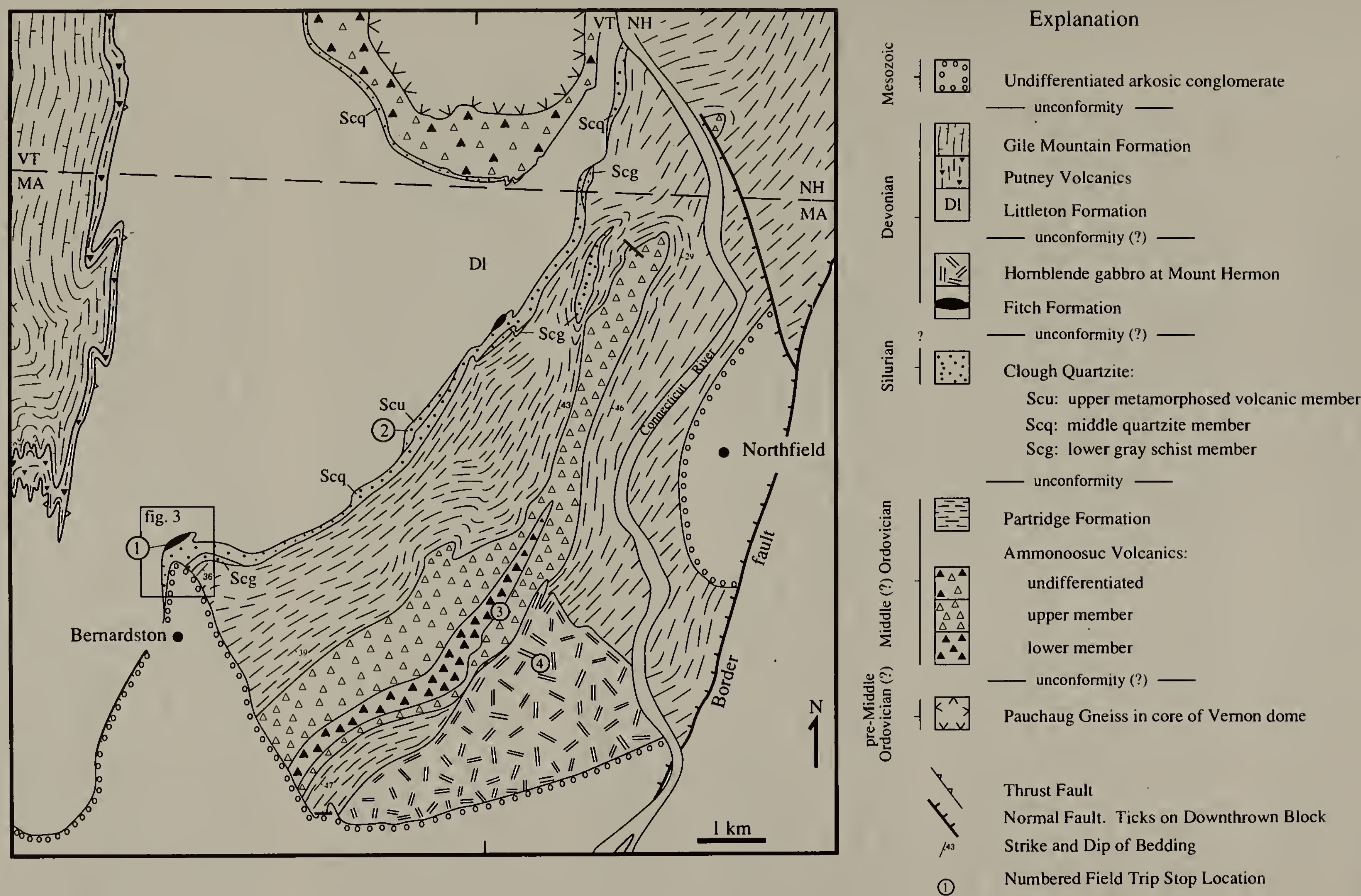



Figure 2. Geologic map of the Bernardston-Northfield area showing locations of field-trip stops 1-4. Distribution of the Gile Mountain Formation and Putney Volcanics from Zen and others (1983) and Hudec (ms). Unpatterned rocks east of the Connecticut Valley Border fault are chiefly Littleton and Erving Formations (Devonian?) (see Zen and others, 1983).

main body of Monson Gneiss (Robinson, 1967; Robinson, Thompson, and Rosenfeld, 1979) and west of the Warwick dome (Schumacher 1981, 1983, 1988).

An inverted section of Silurian Clough Quartzite extends from the Bernardston fossil locality northeastward along the inverted limb of the Bernardston nappe (figs. 2 and 3). I have subdivided the Clough Quartzite in the area into three members. The lower member consists of gray schist which contains a basal polymictic conglomerate. The middle member is made up of metamorphosed orthoquartzite and quartz-pebble conglomerate. The upper member contains feldspathic granulite which probably represents metamorphosed felsic volcanics (Elbert, 1984). Marble and calc-silicate gneiss of the Fitch Formation occur at two locations along the inverted limb (fig. 2).

Figure 2 also shows Silurian strata in a narrow isoclinal syncline I have mapped between the inverted section, along strike from the fossil locality, and the axial trace of the West Northfield anticline (Elbert,1984). I have interpreted this area of Silurian rocks as being on the upper limb of the Bernardston nappe. Reconstruction across the Connecticut Valley border fault to the east indicates that this narrow isoclinal syncline lies on the same axial surface as the Prescott syncline of the Quabbin Reservoir area. This is illustrated in a cross section through the Bernardston-Northfield area and continued eastward across the Connecticut Valley border fault through the Mt. Grace quadrangle (see figure 3 of Robinson and others, this guidebook. In this cross section I have constructed the border fault as a circular arc to allow resetting of the Mesozoic movement and highlight the structural correlations across that fault. An alternate interpretation is that this small syncline is made up of Rangeley Formation rocks in a klippe of the Brennan Hill thrust sheet.

The Mount Hermon Hornblende Gabbro is exposed in the Mt. Hermon pluton in the southern part of the map area (fig. 2). I have mapped this metamorphosed gabbro and numerous related dikes and sills both in the Bernardston-Northfield area and in the Hinsdale-Chesterfield area immediately to the north (Elbert, 1984, 1986). These rocks have intruded strata as young as Late Silurian and were deformed and metamorphosed during the Acadian orogeny. After several seasons of mapping these metamorphosed gabbros I have not found them intruded into the Littleton Formation and suggested in my 1986 NEGSA talk that they are metamorphosed Silurian intrusives.

***Bernardston Fossil Locality***. The importance of this locality is based in the importance of fossil control in all metamorphic terrains. Indeed it has been argued that stratigraphy is the basis of all metamorphic geology (Billings, 1950) and with only a handful of well documented fossil localities in New England, each one provides important stratigraphic constaints for structural and tectonic interpretation.

Fossil remains were first discovered in the marble beds at the Williams' Farm in Bernardston by Edward Hitchcock in 1833 (Hitchcock, 1833, page 295). For historical perspective, Lyell's *Principles of Geology* was published in the early 1830's, while in Devon, T.H. De La Beche, Roderick Murchison, and Adam Sedgwick were formulating ideas which would lead to the establishment of the Devonian System by Sedgwick and Murchison in 1839. Since Hitchcock's discovery the Bernardston fossil locality has been the subject of more than a dozen published contributions and has been visited and studied by some of North America's most prominent geologists. Most recently, Elbert and others (1988) have described an assemblage of Lochkovian (Early Devonian) conodonts from the marbles of the Fitch Formation at this locality.

Hitchcock's (1833) discovery was probably the first report of pre-Carboniferous fossils in New England. Hitchcock later noted (1851) that James Hall had correlated the Bernardston marbles with the Onondaga Limestone (Devonian) of New York based on crinoid fragments. In subsequent publications (Hitchcock and others, 1861) these marbles were referred to as "Upper Helderberg?" (equivalent to "Onondaga" of modern usage). Dana visited the locality several times and concurred with the Devonian age assignment (Dana, 1873, 1877). He also reported the first discovery, by B.K. Emerson in 1877, of fossils in the quartzites immediately above the marbles.

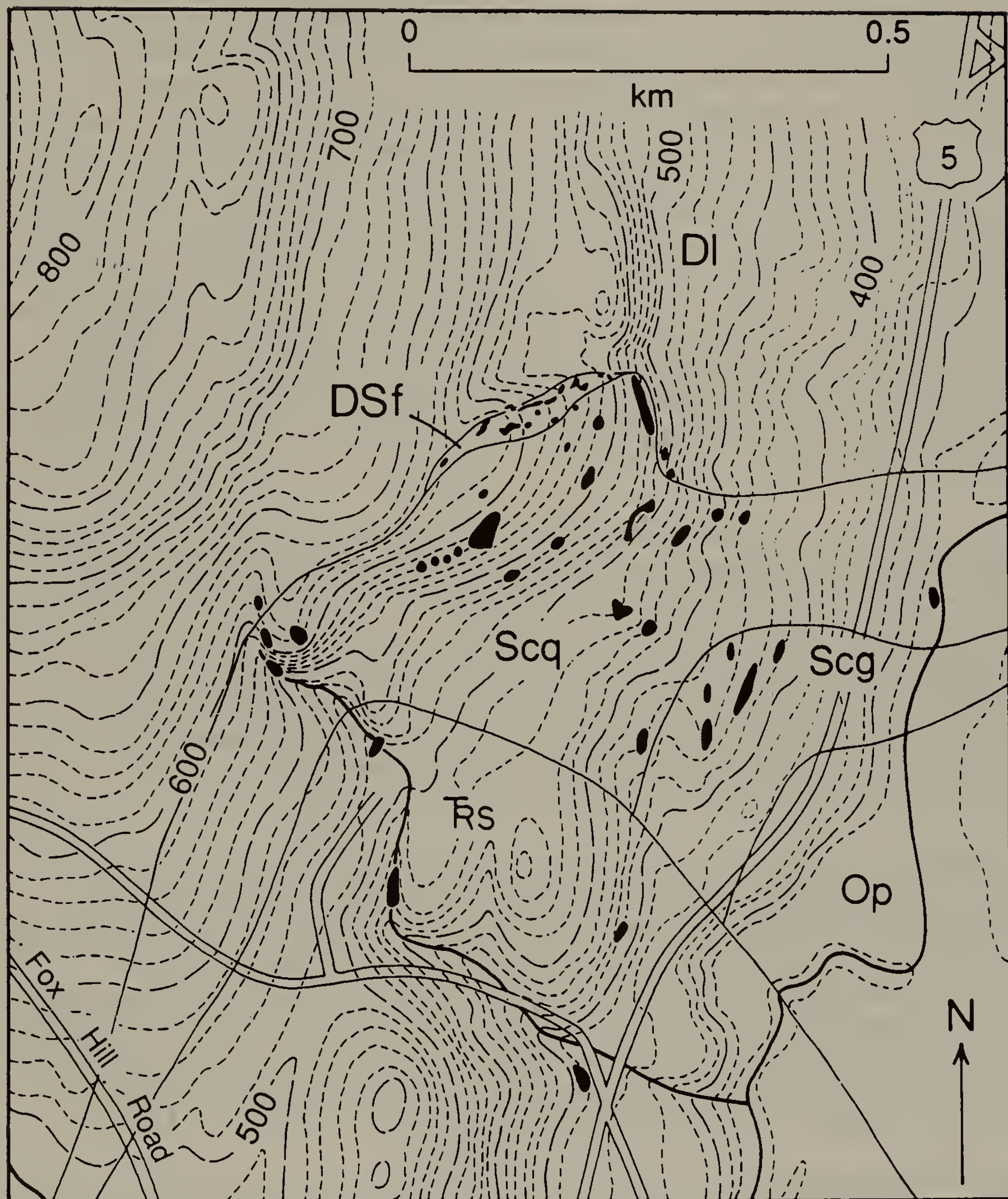



Figure 3. Geologic map (from Elbert and others, 1988) of fossil locality area in the vicinity of Bernardston, Mass. (outcrops shown in black). Trs, Upper Triassic Sugarloaf Formation; Dl, Devonian Littleton Formation; DSf, Lower Devonian and Silurian(?) Fitch Formation; Scq, middle quartzite member of Silurian Clough Quartzite; Scg, lower gray schist member of Clough Quartzite. Contour interval 10 feet; base from U.S. Geological Survey, Bernardston, Mass.-Vt., 7.5-minute quadrangle, 1977.

A more complete study of Dana and Emerson's collections was made by Whitfield (1883), who assigned the marbles to the "Middle Silurian" based on corals. He also reported a brachiopod fauna in the quartzite which he considered "Middle Devonian". Clarke (*in* Emerson, 1898) studied the brachiopods and crinoids in the quartzite and assigned them a Late Devonian age. Schuchert and Longwell (1932) suggested the rocks contained a Late Devonian fauna but did not publish descriptions of the rocks or fossils. Balk (1941) and Cooper and others (1942) confirmed the Devonian age of the strata based on brachiopods reportedly recovered by Balk from the magnetite bed in the marble near its contact with the quartzite. Boucot and others (1958) restudied the specimens of Whitfield, Clarke, and Balk. They demonstrated that the fossils reported by Balk from the magnetite bed were not from Bernardston but probably from Lower Devonian rocks in Nova Scotia. In addition to reexamining the collections of Clarke and Whitfield, Boucot and others (1958) collected new samples of marble from outcrops and calcareous quartzite from the spoil piles of the small, open-pit iron mines at this locality (A.J. Boucot, oral commun., 1983). The marble produced favositids and crinoids, only indicating a post-Early Ordovician age. The calcareous quartzite yielded the brachiopod *Eospirifer* cf. *E. radiatus* (Sowerby) indicating a late Llandoverian to early Ludlovian age. Although the fossiliferous calcareous quartzite bed is no longer exposed, earlier studies clearly established that it was immediately above (stratigraphically below) the marble. Trask's (ms) mapping firmly established the correlation of the quartzite with the Clough Quartzite and the marble with the Fitch Formation even though the marble was included as part of the Clough Quartzite in a subsequent field trip guidebook (Trask and Thompson, 1967).

The recent discovery of what seem to be the world's largest, as well as highest grade (color alteration index = 8), regionally metamorphosed conodonts comprises the most recent installment in the history of this important locale (Elbert and others, 1988). Although conodonts are rare in regionally metamorphosed rocks of lower amphibolite facies, over 1000 recognizable conodonts were recovered from 129 kg of dominantly quartzose marble from pits (figure 4) at this locale. They are all relatively poorly preserved, having reached at least 500 °C. They are generally recrystallized, virtually all incomplete, but few specimens are significantly deformed. The fragmentation of conodont elements in these samples was primarily an effect of deposition in a high-energy environment and not a result of tectonism. The conodonts are indicative of the earliest Devonian *woschmidti* to *eurekaensis* Zones.

The recognition of conodonts at this locality has several implications for the stratigraphy of the Bronson Hill anticlinorium. The conodonts are Lochkovian (earliest Devonian); younger than those reported by Harris and others (1983) from near Littleton, New Hampshire, and extend the regional age range of the Fitch Formation. Figure 5 summarizes the paleontological control on Silurian-Devonian stratigraphy in the Bronson Hill anticlinorium. Boucot and others (1958) have established a Llandoverian to Wenlockian age for the calcareous quartzite which outcropped physically above the marbles Elbert and others (1988) have shown to be Lower Devonian. This is the first paleontological control in adjacent beds to independently support the ~25-year-old structural interpretation of overturned stratigraphy and the presence of regional-scale fold nappes in the Bronson Hill anticlinorium.

**Geology of the Hinsdale Chesterfield Area** (figures 6, 7, and 8): A geologic map of the northern portion of the field-trip area is shown in figure 6. Detailed mapping in the Biscuit Hill area (figure 7 and Stop 6 of this trip) in Hinsdale, NH, delineated a Silurian stratigraphic sequence, all upside down and immediately beneath the Ashuelot pluton (Elbert, 1986). The section is extremely thin; estimated thicknesses of the individual stratigraphic units range from a few inches to ten meters. The characteristic succession of distinctive rock types, coupled with primary facing information from relict graded bedding, has led to the correlation of this stratigraphy with that of the Monadnock area (Elbert, 1986).

This Siluirian stratigraphic sequence, most completely exposed at Biscuit Hill, includes the following: Interbedded gray-weathering and rusty-weathering schist with schist-matrix and quartzite-matrix conglomerates and rare calc-silicate pods of the Rangeley Formation. The stratigraphic top of the Rangeley Formation contains a few thin quartzite beds and grades into the cyclically interbedded quartzites and gray schists of the Perry Mountain Formation. Graded beds in both the Rangeley and Perry Mountain confirm the topping direction. The top of the Perry Mountain contains biotite-rich, massive gray schist with beds and boudins up to 8 meters thick of fine-grained coticule (garnet granulite), coticule conglomerate, and magnetite-grunerite-garnet-apatite-quartz-graphite iron formation. Stratigraphically higher rocks at Biscuit Hill are well bedded, rusty-weathering, sulfidic graphitic calc-silicate granulites and interbedded sulfidic

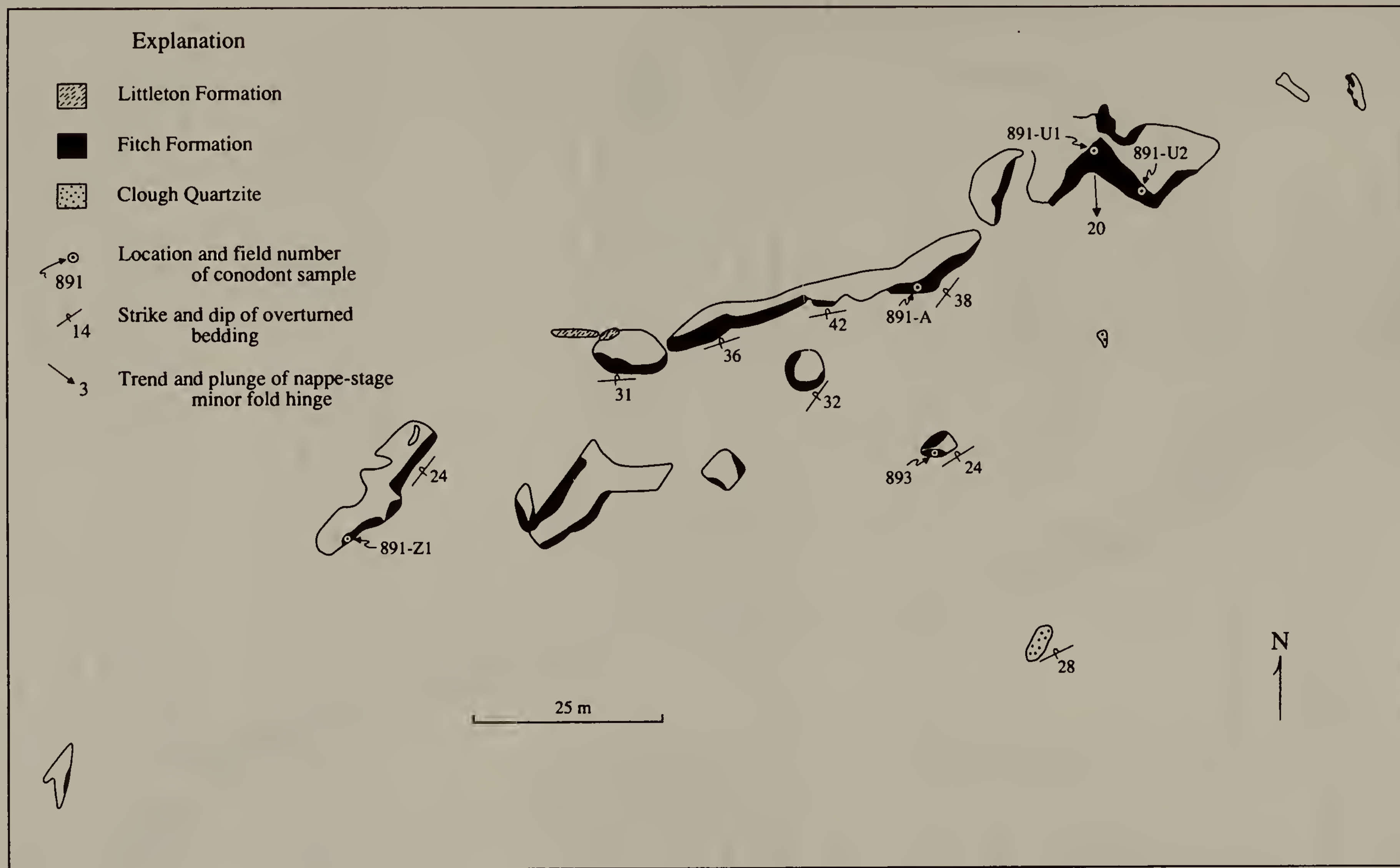



Figure 4. Map (from Elbert and others, 1988) showing conodont-sample localities and stratigraphic units exposed (outcrops patterned) in pits (outlined) about 1.5 km NNW of Bernardston, Mass. (see figs. 1-3 for regional and local geologic setting). Sample location 893 lies in pit a few meters southwest of right-angle bend in logging road. Other marble-bearing pits are north of road, mostly under the large hemlocks. Structure symbols apply to closest outcrops.

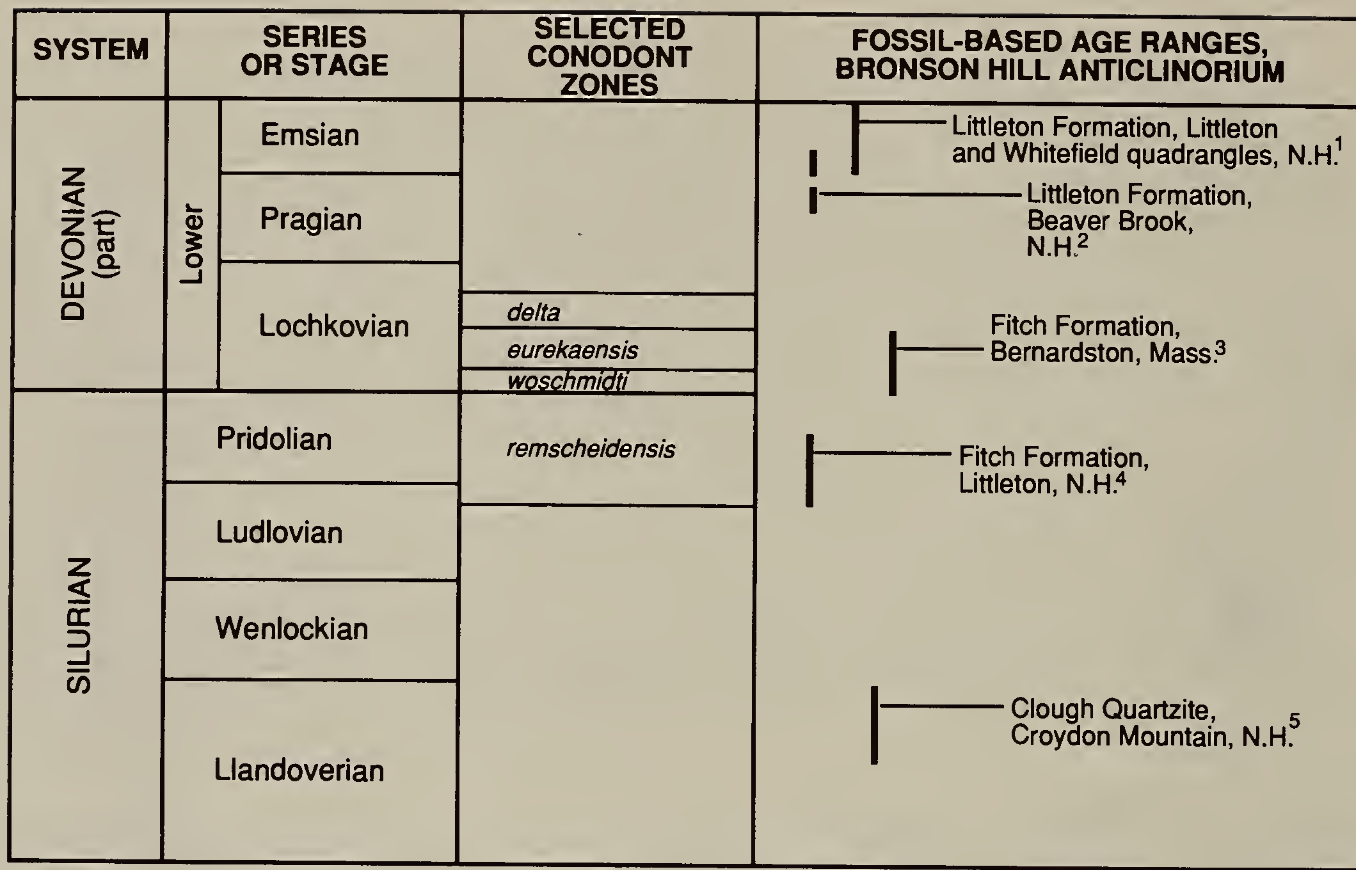



Figure 5. Fossil-based age ranges of stratigraphic units in the Bronson Hill anticlinorium.
1. *Amphigenia*, *Eodevoniaria*, and *Euryspirifer* (brachiopods) from Pageau Farm, Mormon Hill, and Dalton Mountain localities, New Hampshire (Boucot and Arndt, 1960); 2. *Leptocoelia* (brachiopods) from Beaver Brook locality, New Hampshire (Boucot and Rumble, 1980); 3. *woschimidti* Zone conodonts from the Fitch Formation at Bernardston (Elbert and others, 1988); 4. lower *remscheidensis* Zone conodonts from the lower and middle parts of the Fitch Formation in the Littleton and Lower Waterford 7.5-minute quadrangles (Harris and others, 1983); 5. faunas from Clough Quartzite at Bernardston, Mass., Croydon Mountain, New Hampshire, and Skitchewaug Mountain, Vermont (Boucot and Thompson, 1963; Boucot and others, 1958).

Mg-biotite schists of the Francestown Formation, stratigraphically overlain by well bedded, clean, actinolite-garnet-calcite calc-silicate gneisses, garnet granulites and interbedded purplish biotite granulites of the Warner Formation. The lower part of the Warner is generally richer in calc-silicate rocks and the upper part in biotite granulites. Numerous metamorphosed gabbro sills intrude the strata and are presumed to be related to the Mt. Hermon Hornblende Gabbro and the associated dikes and sills in the the Bernardston-Northfield area.

The distinctive Perry Mountain iron formation rocks are particularly important to both local mapping and regional interpretations (Elbert, 1986; Thompson and others, 1987 and in preparation). The magnetite-garnet iron formation and coticule have been correlated with strata which had previously been assigned to the Littleton Formation in the Mt. Grace quadrangle (Robinson, 1967; Huntington, 1975), just above the root zone of the Bernardston nappe (Thompson and others, 1987; Robinson, 1988; Robinson and others, trip C-4 of this volume). This correlation of such a thin lithogenetic unit may seem somewhat long range. When the Mesozoic listric motion on the Connecticut Valley border fault is restored, however, the two locations are only a few kilometers apart and at a comparable structural level.

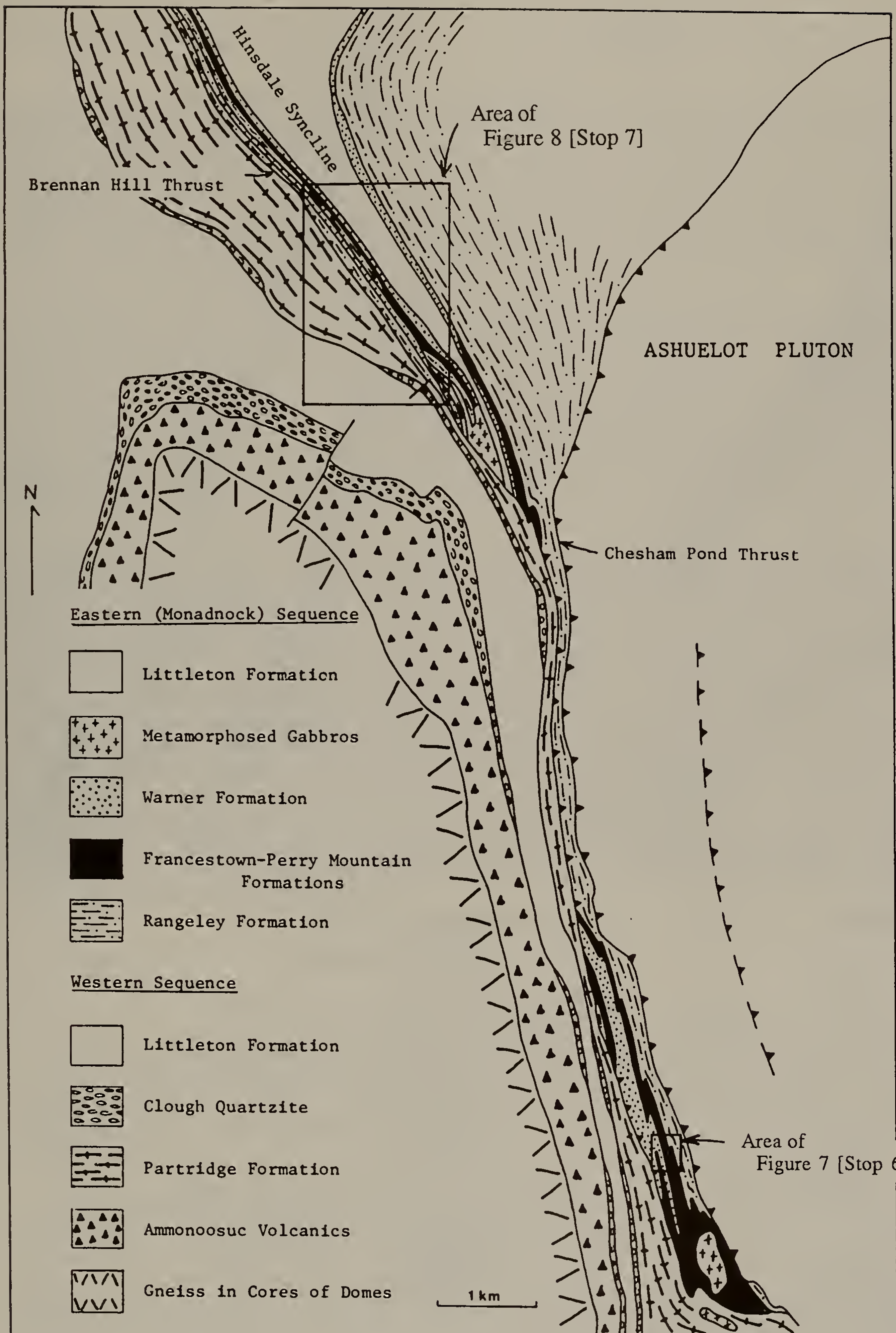
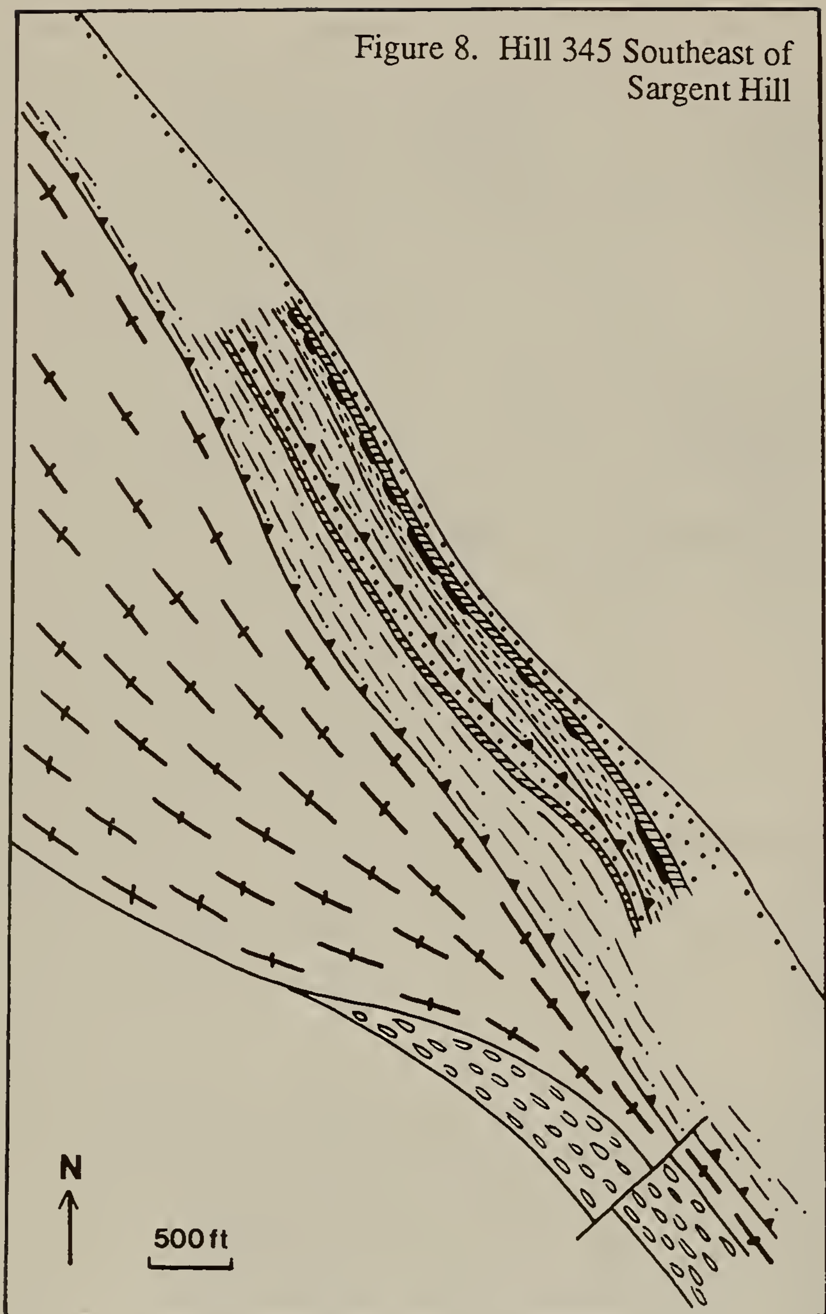
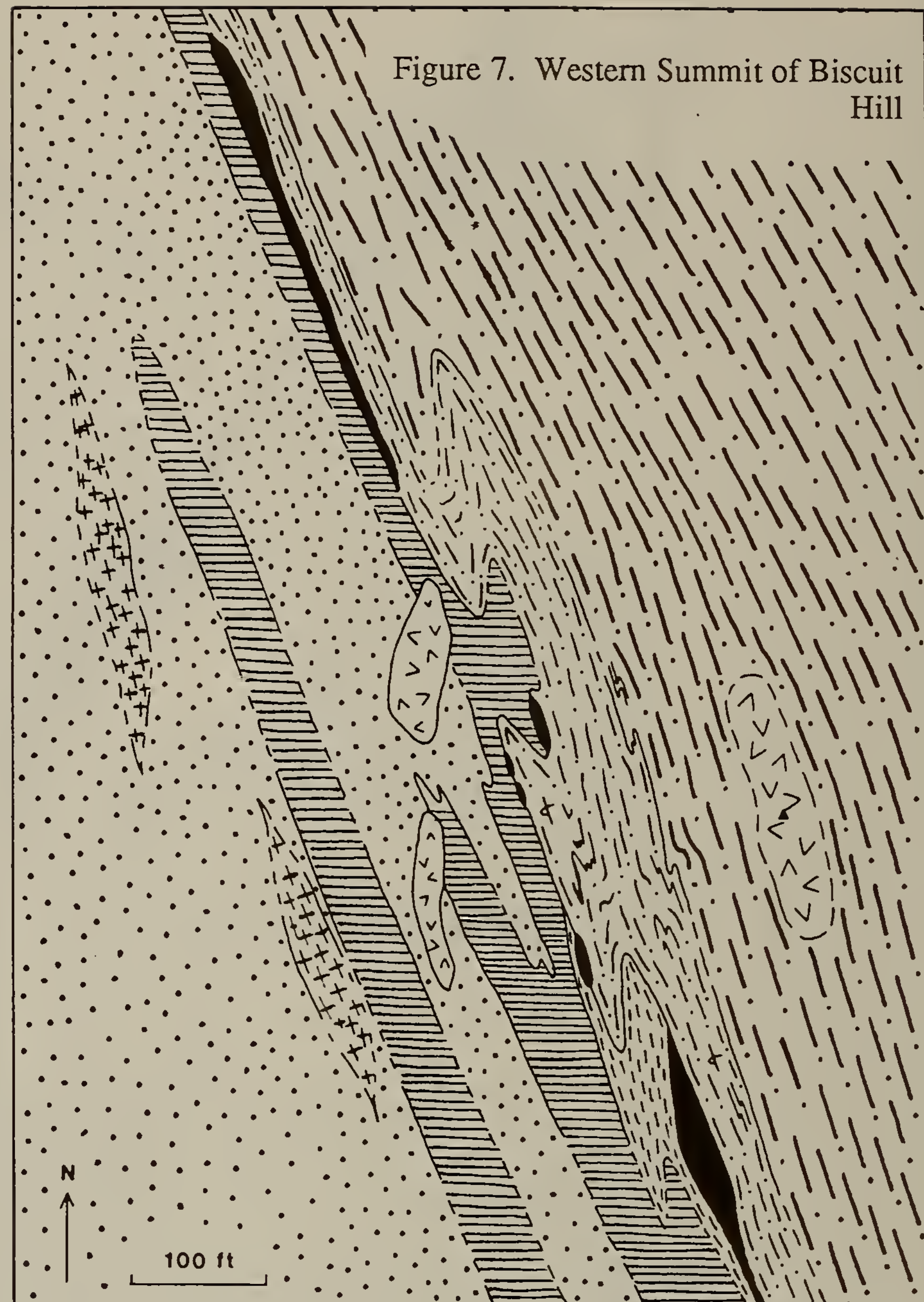



Figure 6. Generalized geologic map of the Hinsdale-Chesterfield area showing the Hinsdale syncline of Monadnock sequence stratigraphy in the hanging wall of the Brennan Hill thrust (from Elbert, 1986). Dashed line with teeth in Ashuelot pluton is shear zone mapped by Armstrong (1986).




Western Sequence

Lower Devonian — Littleton Formation

Silurian — Clough Quartzite

Middle Ordovician — Partridge Formation

Eastern (Monadnock) Sequence

Lower Devonian — Pegmatite; Littleton Formation

Silurian? — Metamorphosed Gabbros

Silurian — Warner Formation; Francestown Formation; Perry Mountain Formation (8, 7): massive gray schist, magnetite-garnet iron formation, interbedded quartzite and gray schist; Rangeley Formation

Figure 7. Detailed geologic map of a portion of Biscuit Hill. East-dipping, inverted, multiply folded Monadnock sequence faces down (west) toward Brennan Hill thrust. Map-scale folds formed during the backfold and dome stages (from Elbert, 1986).

Figure 8. Geologic map of region southeast of Sargent Hill. Imbricated right-side-up Silurian part of Monadnock sequence, previously mapped as Clough and Fitch by Trask (1964) and Hepburn and others (1984), lies in thrust contact with Partridge Formation in the core of the Bernardston nappe (Elbert, 1986).

Four phases of Acadian deformation have been found in the rocks in the Hinsdale-Chesterfield area. Westward-overturned, isoclinal folding first produced the overall inversion of the sequence at Biscuit Hill, where rocks are on the upper limb of the Hinsdale syncline (figure 6). Isoclinal folding in the western-sequence rocks, producing the Bernardston nappe, is postulated to also have occured at this time. This folding was followed by thrust faulting, which juxtaposed upside-down eastern-sequence rocks against the western-sequence strata. Two later phases of deformation have been documented in outcrop-scale folds, rock fabrics, and map pattern, and correspond to the backfold and dome stages of regional deformation.

One of the interesting aspects of the structural history of the region is the lack of a recognizable fabric associated with the thrust faults. In fact thrust faulting in the area is only recognized on the basis of stratigraphic interpretation and mapping. The identification of the west-facing package of rocks believed to have been originally deposited east of their present position in contact with the Middle Ordovician Partridge Formation, in the core of the Bernardston nappe, is taken as the evidence for west-directed thrusting. The stratigraphic character of the rocks on either side of the thrust, the regional tectonic level, and the westerly transport direction of the thrust coupled with the assumption that thrusts propogate tectonically upwards suggest that this thrust correlates with the Brennan Hill thrust defined by Peter Thompson in the Monadnock quadrangle (Thompson, 1984, 1985).

Mapping northward into Chesterfield has yielded reinterpretation of the portion of the Bronson Hill anticlinorium covered by the recently published Bedrock Geology of the Brattleboro Quadrangle, Vermont-New Hampshire (Hepburn and others, 1984; see also their field-trip guide in this volume). In this part of the area an extensive upright sequence of rocks has been delineated, which, with the inverted section mapped northwards from Biscuit Hill, indicates the presence of a major pre-thrust recumbent syncline (figures 6 and 9) I have termed the Hinsdale syncline (Elbert, 1986). Also in this part of the area, the Brennan Hill thrust has cut down to within 30 meters of the Clough Quartzite which is on the overturned limb of the Bernardston nappe and thus probably cut the anticlinal axial surface of the nappe itself. Several imbricate splays of the thrust, some involving Partridge Formation rocks as well as the Silurian rocks, have been identified (figure 8). One of these imbrications was traced for over three-quarters of a mile and will be visited on the last stop of this field trip. Detailed mapping of these splays is not routinely possible due to incomplete exposure and the preponderance of pegmatite and gabbro dikes and sills. These intrusions obscure contacts of stratified rocks and cut out an indeterminable amount of the stratigraphy in a region where entire thrust imbrications can be demonstrated to be no more than ten meters thick and contain stratigraphic units less than a meter thick.

## METAMORPHISM

The complex tectonic development of the Bronson Hill anticlinorium and Merrimack synclinorium in central Massachusetts and southern New Hampshire has resulted in an equally rich metamorphic history. Figure 10 is a map of composite metamorphic zones in the region compiled from Robinson (1986), Robinson and others (1978), Tracy (1978), Tracy and others (1976), Robinson and others (1982), and my own mapping.

The region of figure 10 contains several ages of metamorphism. Relics of a high-pressure, sillimanite-orthoclase grade metamorphism that predates the deposition of Silurian-Devonian strata in the region and was substantially reworked by later events have been studied by Roll (1987) in the Pelham dome. The present pattern of metamorphic zones (figure 10) crosses stratigraphic boundaries and includes the Silurian-Devonian rocks of the region. This metamorphism has been considered to be largely Acadian in age (Robinson, 1986; Zen and others, 1983), however, a recent Rb-Sr mineral isochron from the middle-Ordovician cover rocks of the Keene dome reveals a final isotopic equlibration during the Alleghenian (Gromet, 1988 and personal communication, 1988) and suggests that there may be a more significant post-Acadian metamorphic history than previously recognized.

In the Bernardston-Chesterfield area the metamorphism peaked at considerably lower grade than in the central Massachusetts high-grade zone (figure 10). Rocks in the region of this field trip range from the garnet zone, at the fossil locality (Stop 1), to the sillimanite-staurolite zone, at Biscuit Hill (Stop 6).

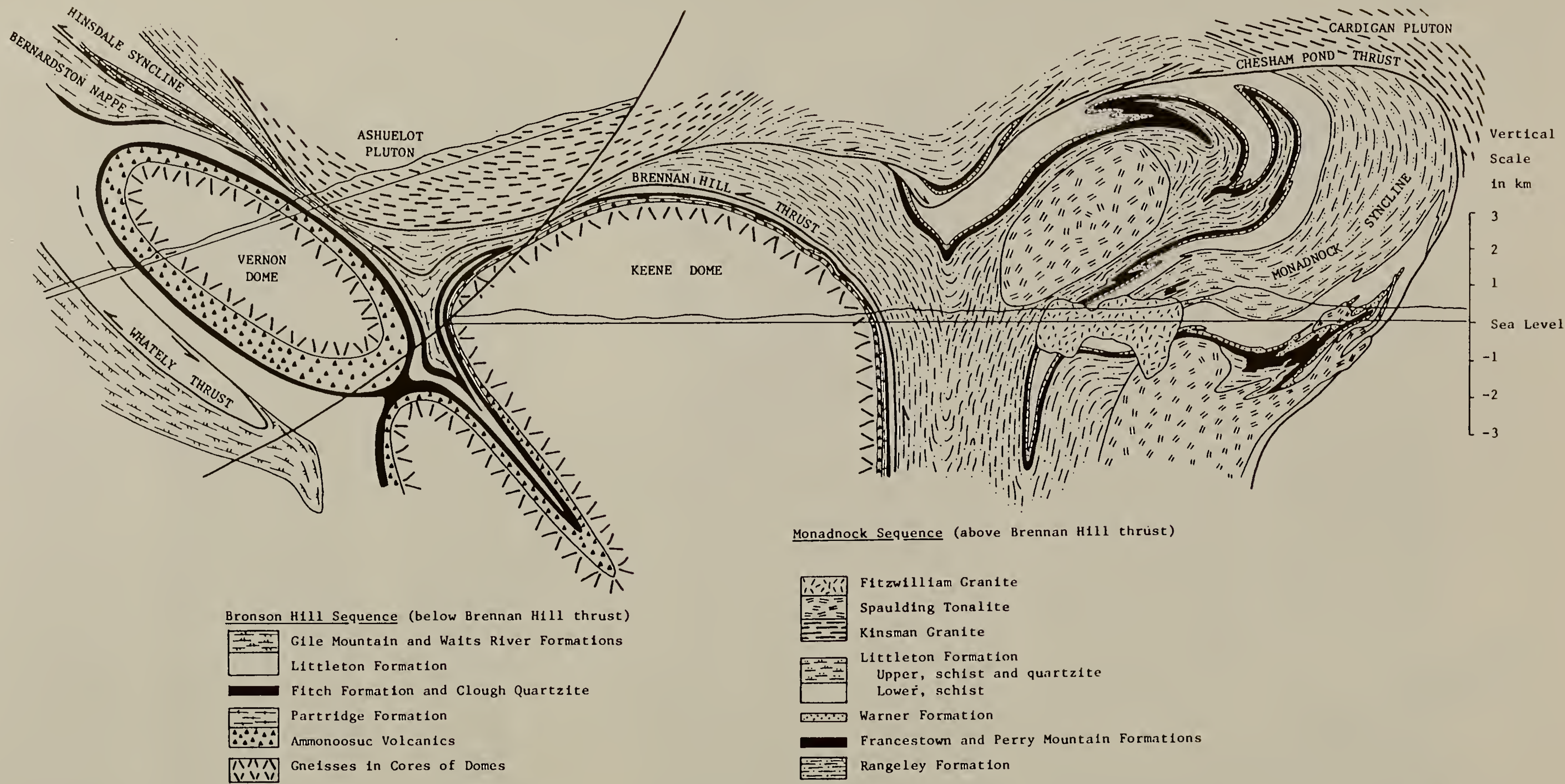



Figure 9. Generalized structure section across the Hinsdale and Monadnock areas based on work of Elbert (1986) and Thompson (1985) as well as older data. Location shown by east-west line in southern New Hampshire in figure 2 of Robinson and others, this volume. Resetting across the Mesozoic Connecticut Valley border fault is approximated for conceptual consistency and ignores the known movement vector transverse to the the line of section. Out-of-section movement vectors also complicate the Chesham Pond fault zone. Details of surface geology east of Chesham Pond fault are omitted.

Metamorphic Zones and Features:

- VI Garnet-Cordierite-Sillimanite-K-feldspar
- V Sillimanite-K-feldspar
- IV Sillimanite-Muscovite-K-feldspar
- III Sillimanite-Muscovite
- II Sillimanite-Staurolite
- $I_k$ Kyanite-Staurolite
- $I_a$ Andalusite-Staurolite
- G Garnet
- Bio Biotite
- C Chlorite
- T-J Mesozoic Sedimentary and Volcanic Rocks
- ○ Pre-Acadian Metamorphism (Roll, 1987)
- ◆ Sillimanite Pseudomorphs after Andalusite

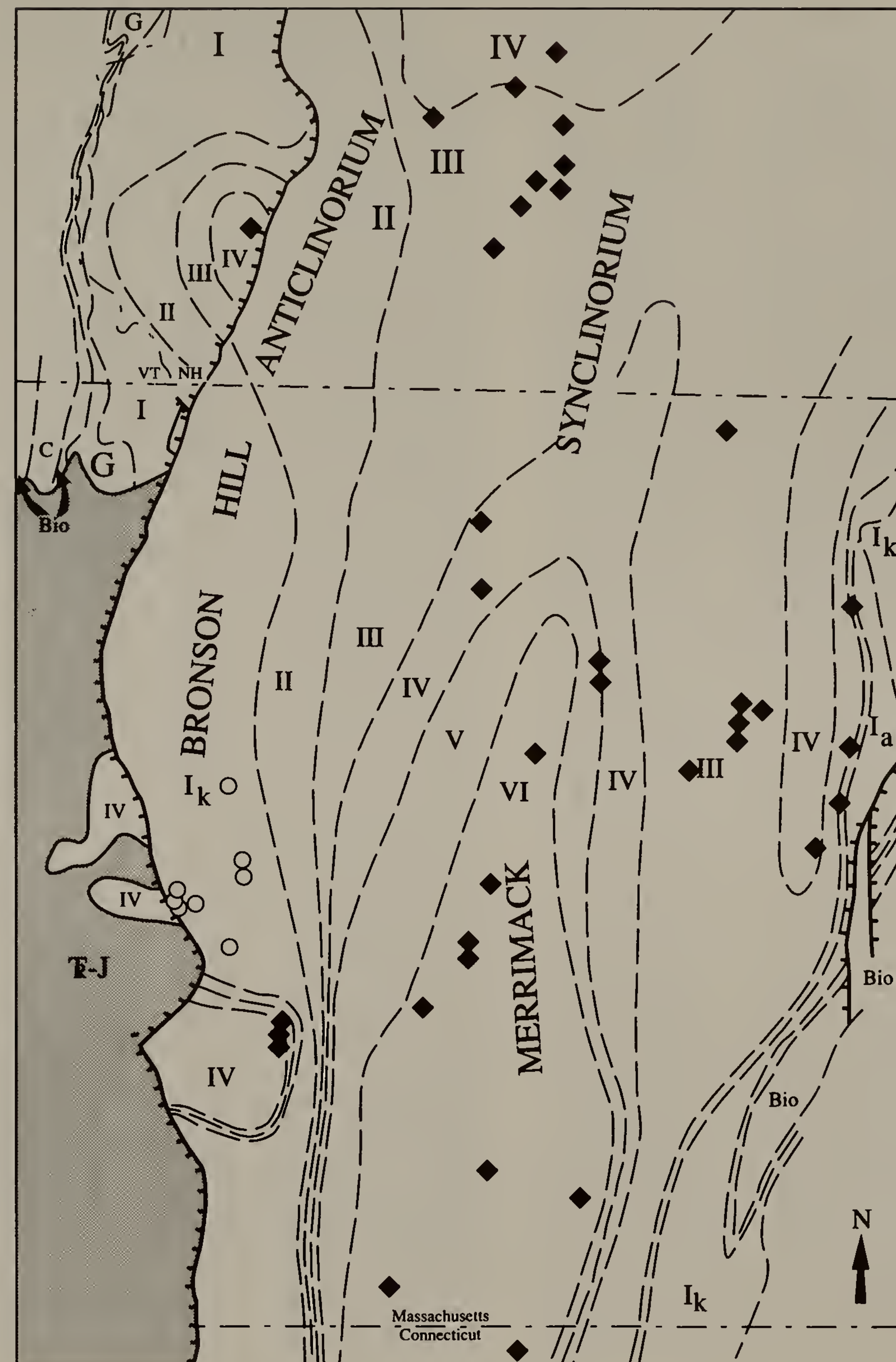



Figure 10. Generalized map of metamorphic zones and features in west-central Massachusetts and adjacent states.

These rocks represent structural levels ~5 km higher than rocks now exposed just to the east, across the Connecticut Valley border fault. Within the Bernardston-Chesterfield area the transition from lower- to higher-grade rocks at successively higher structural levels (compare figure 10, this chapter, and figure 2 of Robinson and others, this volume) is a classic example of an inverted metamorphic gradient. Indeed the metamorphic sequence along most of the western margin of the Bronson Hill was recognized as inverted by Thompson and others (1968). Throughout the region isograds are not spatially related to the gneiss domes of the region, as they are in the adjacent Connecticut Valley-Gaspé synclinorium. Chamberlain and Lyons (1983) found the isograds do not have a simple spatial correlation to syntectonic plutons (an exception is the pattern of metamorphic zones around the Belchertown pluton in the southwestern part of figure 10). Thompson and others (1968) suggested that the early, west-directed, regional fold nappes carried hotter rocks over colder ones and established the isogradic inversion. The new structural and petrologic interpretations and correlations in the Bernardston-Chesterfield region (Elbert, 1984, 1986, 1987; Thompson and others, 1987) suggest instead that early control on metamorphism came from emplacement of the hot, west-directed, thrust sheets that post-dated the fold nappes, with final isograd geometry controlled by isotherm relaxation prior to regional backfolding and gneiss-dome formation.

Sillimanite pseudomorphs after andalusite ("andalumps") are present in a wide region of the Merrimack belt and a restricted region of the Bronson Hill belt (figure 10). In the Bernardston-Chesterfield area "andalumps" are present above the Chesham Pond thrust in the pelitic rocks structurally above the Ashuelot pluton (compare figure 10 and figure 6). These pseudomorphs suggest the rocks in the highest tectonic level in the field area followed a grossly counter-clockwise path through P-T space similar to that described by Spear and others (1983) in rocks believed to be from the same tectonic level in the Bellows Falls area of New Hampshire and Vermont.

In the Hinsdale area, New Hampshire, it has long been known that the sillimanite isograd is roughly concentric to the Ashuelot pluton of Kinsman Granite (Moore, 1949; Trask, 1964; Thompson and others, 1968). In an earlier tectonic model (Thompson and others, 1968) this was thought to be due to nappe-stage recumbent folding of the Kinsman, possibly still in a partially liquid state, to a position tectonicallly above the Bernardston nappe and the parautochthonous Vernon dome. In the recent model (Elbert, 1986; Thompson, 1985; Robinson, 1987; Thompson and others, 1987) the Kinsman of the Ashuelot pluton is considered to have been emplaced largely in a solid state by the Chesham Pond thrust, tectonically above the Brennan Hill thrust sheet.

The sillimanite-in isograd in the field-trip region roughly corresponds to the beginning of the zone of staurolite breakdown. Within the staurolite-breakdown zone abundant pseudomorphs of staurolite are present. These pseudomorphs are largely made up of muscovite, garnet, quartz, sillimanite, plagioclase, graphite, ilmenite, and biotite with various amounts of relict staurolite in the their cores. The pseudomorphs are flattened and sheared into the dominant foliation in the rocks and formed during the prograde breakdown of staurolite. These prograde pseudomorphs are found in the Partridge, Rangeley, and Littleton Formations and are similar to some of the pseudomorphs studied by Guidotti (1968) and Foster (1983) from the Rangeley Formation in Maine. These pseudomorphs are different from the retrograde chlorite-bearing pseudomorphs after staurolite that are abundant near the north end of the Vernon dome and present at stop 6 of the last NEIGC trip to Bernardston-Chesterfield area led by Trask and Thompson (1967).

My own detailed petrologic work has concentrated on the region of the sillimanite-in isograd and the staurolite-breakdown zone. Within this region parts of the reaction histories of the rocks are recorded in zoned garnets which, in a few rare instances, contain important inclusions of plagioclase and chlorite. I have obtained mineral analyses and detailed garnet zoning data from over thirty thin sections and more than a hundred garnets. The data include WDS spot analyses, zoning traverses, and backscattered-electron images all done on the JEOL 733 microprobe at Rensselaer Polytechnic Institute, Troy, NY.

Chemical zoning in garnet has been described by innumerable workers throughout New England and the world and has long been recognized as a potentially powerful key to metamorphic histories (Tracy, Robinson, and Thompson, 1976; Thompson, Tracy, Lyttle, and Thompson, 1977; Spear and Selverstone, 1983; Spear and others, 1983; Tracy, 1982; Karabinos, 1984; Chamberlain, 1986). Garnet zoning in the high-grade rocks of central Massachusetts has been divided into three distinct types (Tracy and others, 1976; Robinson and others, 1982). These include: garnets that have grown during prograde metamorphism by the continuous addition of new material to their rims while the diffusion rate within the garnet was too

slow to permit continuous equilibration of the interior and the exterior of the garnet; garnets that were homogenized by diffusion when they reached sufficiently high grade and now only record zoning at their rims and only related to retrograde reactions; and garnets that have homogenized by diffusion and have been later modified by localized, retrograde ion-exchange reactions where the garnet is in contact with another Fe-Mg-bearing phase. This last type of garnet presumably escaped pervasive ion-exchange re-equilibration due to the absence of a fluid to serve as a diffusional pathway beyond the region of actual contact with the other Fe-Mg phase. The zoned garnets from the Bernardston-Chesterfield area are distinct from those recognized by Robinson and others (1982) elsewhere in the region.

Zoned garnets from the lower sillimanite zone in the Bernardston-Chesterfield area can be sorted out by textural and chemical criteria which exactly correlate to and have been determined by tectonic level (Elbert, 1987). At Biscuit Hill (Stop 6) pelitic rocks above and below the Brennan Hill thrust contain virtually identical assemblages and staurolite pseudomorphs. Garnet zoning in these rocks, however, is distinct across this tectonic boundary.

Garnet below the Brennan Hill thrust, in the Partridge Formation, are subhedral to euhedral and only contain tiny (< 1 μm across) inclusions in a crystallographically controlled, radial pattern within their core regions. These garnets contain three different chemical zones. The zoning is concentric and readily viewed in microprobe traverses (figure 11A). The cores of these garnets are unzoned and euhedral. These cores correspond to the region containing the fine, radial inclusions. Surrounding the unzoned cores is a broad (300-400 μm in many of these garnets) region dominated by decreasing spessartine and grossular, increasing almandine and pyrope, and constant Fe/(Fe+Mg) (=0.90) from core towards rim. This broad intermediate region is surrounded by a thin (typically 40-50 μm) rim marked by large increases in almandine, grossular, and Fe/(Fe+Mg), with large decreases in spessartine and pyrope.

Garnet porphyroblasts from the Rangeley Formation schists above the Brennan Hill thrust are distinct from those below the thrust. These hanging-wall garnets also contain unzoned euhedral cores (figure 11B). These cores are surrounded by an intermediate zone rich in large (up to 0.1 mm across) inclusions of quartz, ilmenite, and, in a few instances, plagioclase. The inclusions are the remains of a tectonic fabric that predates the backfold- and dome-stage fabrics present in the matrix of the rock. This included fabric may be related to the pre-thrust folding. The intermediate zone in these hanging-wall garnets contains similar overall zoning trends to those in the footwall garnets. In detail, however, the intermediate regions in the hanging-wall garnets are different. Their overall zoning trends are punctuated by several peaks high in spessartine, and others high in grossular. Backscattered-electron images reveal that some zones only partially surround a given garnet and are truncated by a later and locally "deeper" resorption zone forming complex compositional "unconformities." Chemical trends in the thin rims of the garnets above the Brennan Hill thrust are identical to those below the thrust.

The unzoned cores in both these garnet types may represent garnet growth during effective thermodynamic invariance. While this interpretation is straight forward, it seems problematic in light of even optimistic estimates of the rocks' assemblage history. A hypothesis I prefer is that these garnets were quenched early in the process of diffusional homogenization. Several authors have noted the transition from zoned to unzoned garnets in the lower to middle sillimanite zone zone and have attributed this transition to the onset of rapid diffusion over a narrow temperature interval. The exponential relationship of simple diffusion-rate laws has allowed petrologists to consider this transition to be instantaneous to a first approximation (e.g. Yardley, 1977; Anderson and Olimpio, 1977). In these rocks, however, diffusion may have been selectively enhanced in the core region by the oriented grain boundaries of the tiny radial inclusions. In fact, factors such as initial zoning profile, heating rate, fluid composition, inclusion size and distribution, and garnet size and spacing may all affect the diffusional process. These factors combined with the possibility of "tectonic quenching" during either the backfold- or dome-stage deformations add to the attractiveness of nascent diffusional homogenization as the source of the unzoned cores.

Zoning in the broad intermediate regions in the footwall garnets has been interpreted (Elbert, 1987) as recording garnet growth, during staurolite breakdown, along the continuous Fe-Mg-Mn-Ca-K-Na reaction:

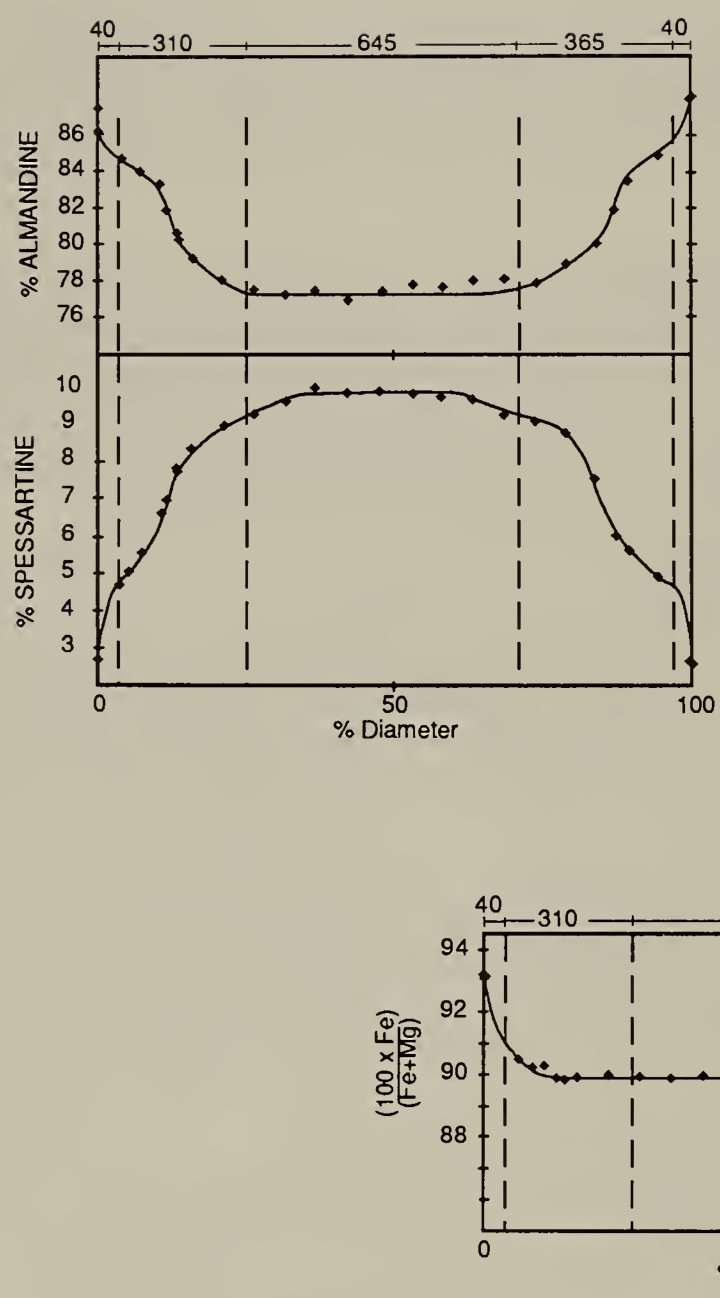



Figure 11A. Microprobe traverse from a garnet porphyroblast in the Partridge Formation, below the Brennan Hill thrust. Zoning is concentric and continuous, typical of footwall garnets.

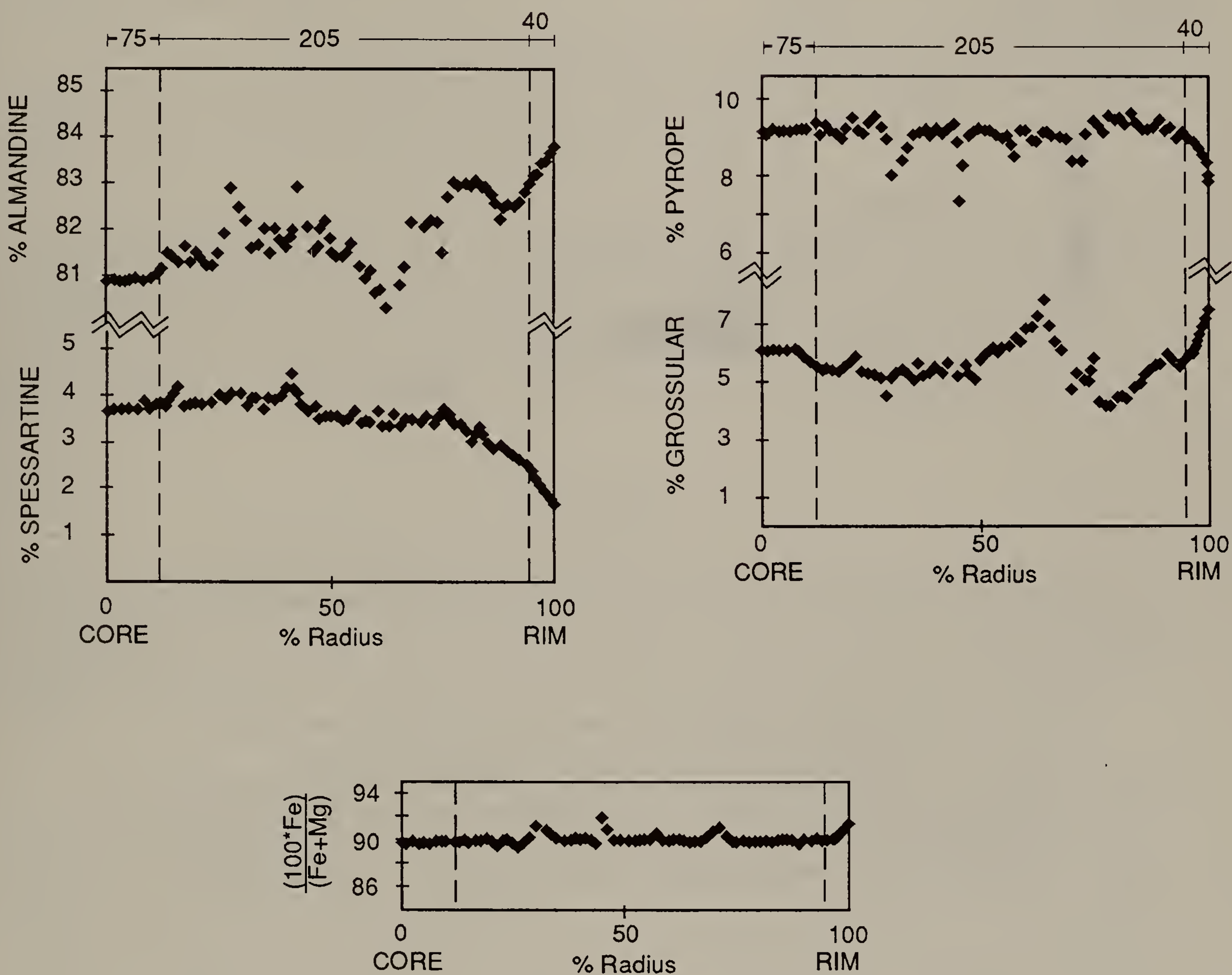



Figure 11B. Microprobe half-traverse from a garnet porphyroblast in the Rangeley Formation, above the Brennan Hill thrust, Biscuit Hill, Hinsdale, N.H. The traverse shows zoning complexity typical of garnets above the thrust.

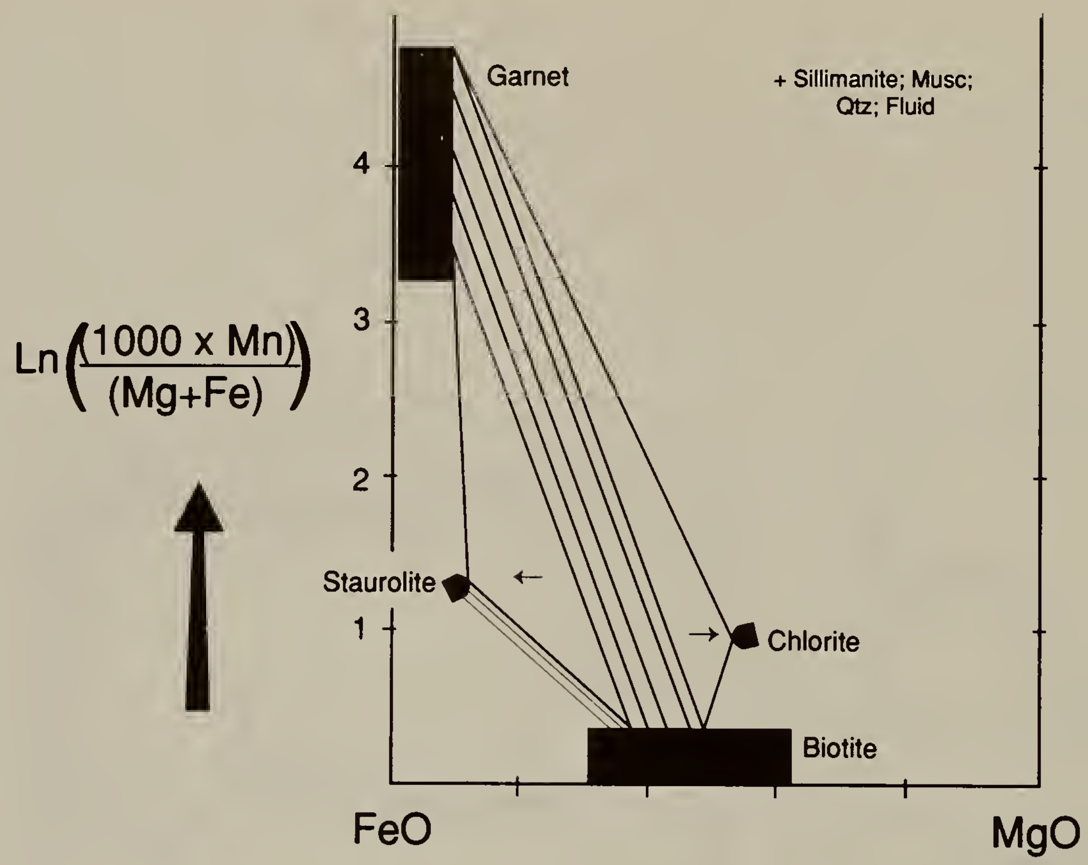



Figure 12. Projection from sillimanite, muscovite, quartz, and fluid onto the plane MnO-FeO-MgO with the MnO vertex removed to infinity and scaled logarithmically to supress the plotting position of high-Mn phases (in this case garnet) while expanding the scale in the region of low-Mn phases. The continuous staurolite-breakdown reaction of Elbert (1987), in the MnKFMASH system, in represented be the three-phase triangle involving staurolite, garnet, and biotite.

Fe,Mg,Mn-Staur + Qtz + Na,K-Musc + Plag $\rightarrow$ Sill + Fe,Mg,Mn-Bio +
Fe,Mg,Mn,Ca-Gar + K-richer-Musc + Ab-richer-Plag +$H_2 0$

This reaction is graphically depicted in figure 12. Forward modelling of garnet growth and the discovery of chlorite inclusions in some garnets suggests that reactions involving chlorite may also be important.

The thin rims on garnets above and below the Brennan Hill thrust are consistent with garnet growth, during cooling, of the assemblage garnet - biotite - sillimanite - muscovite - quartz - fluid once the system is effectively depleted in staurolite and manganese component.

Although the broad intermediate zones in the hanging-wall garnets probably grew along the same equilibrium as those in the footwall garnets, the discontinuous zones rich in spessartine record periods of partial garnet resorption not found below the thrust. These successive periods of garnet growth and resorption make modelling of related reaction histories particularly tricky. However, reasonable interpretations of reaction histories utilizing the Gibbs' method approach (Spear and others, 1982; Spear and Selverstone, 1983) supports the suspicion that the high-grossular zones correspond to garnet growth under temporarily increased pressure.

Partial pressure-temperature histories in these rocks have been calculated. Although garnets above and below the Brennan Hill thrust record broadly similar reaction histories, lack of detailed correlation of zoning in the intermediate regions of these garnets suggests that the P-T paths of the hanging-wall and footwall rocks were different for all but the last 50-100 µm of garnet growth.

An interesting aspect of the hypothesized diffusional homogenization of the garnet cores is the affect of chemical exchange with the matrix. Exchange between the core of a zoned mineral and its matrix results in metasomatism of the reactive composition of the rock. This effectively increases the variance of the assemblage and has the potential to drive and control subsequent reactions. Of course it is difficult to determine whether these garnets were beginning to homogenize during the later stages of or after their growth. The presence of zoning at their rims suggests that final growth was either before the onset of significant diffusion in these garnets or after they had cooled back down to low-diffusivity temperatures.

## SUMMARY

Monadnock-western Maine Silurian rocks have been recognized in the Bernardston-Chesterfield area. This eastern stratigraphic sequence has been isoclinally folded and then thrusted westward onto Bronson Hill-sequence rocks which make up the Bernardston fold nappe. The rocks were then folded again during the regional backfold stage of deformation. The final Acadian phase of folding locally involved the formation of the Vernon gneiss dome, many map- and outcrop-scale folds, and a pervasive southeast plunging lineation. Mesozoic faults and joints comprise the most recent structures in the region.

The sillimanite-in isograd is broadly parallel to the trace of the Chesham Pond thrust. This suggests a genetic link between thrusting and metamorphism. In detail, however, the isograd cross-cuts all lower tectonic levels indicating that final isograd geometry post-dates thrust faulting. Sillimanite in staurolite pseudomorphs is deformed by small-scale backfold- and dome-stage structures and isograds have been folded during the dome stage. Chemically zoned garnets above and below the Brennan Hill thrust record different pressure-temperature histories in their intermediate regions. Core-region records have been obliterated by diffusion. Thin growth rims on garnets are identical across the fault. These relationships suggest that while the P-T paths of the rocks straddling the Brennan Hill thrust converged before isograd development, the convergence was fairly late in the garnet-growth history while relatively early in the the rich tectonic history.

## ACKNOWLEDGMENTS

I am grateful to many friends associated with the University of Massachusetts over the past several years. I wish to thank Peter Robinson for his time, interest, and help with all aspects of this project. Peter Thompson, Kurt Hollocher, Spike Berry, and John Schumacher have been important resources on regional geology, geochemistry, and petrology. I am particularly indebted to Spike for his timely and helpful review of the manuscript. Tom Armstrong was an especially important part of this work acting far beyond his appointed position of field assistant. Tom's enthusiasm to be more than the other end of the tape measure during the detailed work on Biscuit Hill was critical. As always, Julie and Sarah Elbert have provided important support and field assistance.

Microprobe work supporting the petrologic interpretations was done at RPI and I thank Frank Spear, Bruce Watson, and Dave Wark for making that facility available. I especially thank Karleen Davis, now of

Link Analytical, for making the RPI probe so accessible and the hours spent there so inviting, and informative. Anita Harris and Kirk Denkler added valuable comments in the field during a visit to the fossil locality. I am obliged to Wally Bothner for being a patient and sane guidebook editor.

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Finally I would like to dedicate this chapter to Leo Hall. His caring and patience in the field made him an inspirational teacher. His guidance is reassuring every time I bend down to take a closer look.

## ITINERARY

Starting Point is Streeter's Barber Shop, on Route 10 immediately west of the intersection with Interstate 91 (exit 28) in Bernardston, Massachusetts. All the stops on this trip are on private property. Please respect both the outcrops and the rights of the landowners. Those following this guide in the future are responsible for securing permission before entering private property.

| Mileage | Description |
|---|---|
| 0.0 | Proceed west on Route 10 to the intersection with Route 5. |
| 0.1 | Bernardston Auto Exchange-Streeter's General Store on left. |
| 0.5 | Stop sign, junction Route 5. Turn right and proceed north on Route 5. |
| 0.6 | Fox Hill Road on left, continue north on Route 5. |
| 0.9 | Outcrop on left of conglomeratic Upper Triassic Sugarloaf Formation at the northern extreme of the Deerfield basin. Continue northward on Route 5. |
| 1.1 | Pull off road onto wide grassy shoulder on right (east) side of road and park. Cross road (CAREFUL...Route 5 is heavily travelled at times). Walk up short dirt track, through gate into pasture. CLOSE GATE BEHIND YOU!! |

**STOP 1. Bernardston Fossil Locality** (Figures 3, 4, and 5; approximately $1\frac{1}{2}$ hours)

Outcrops in the lowest part of the pasture are part of the lower member of the Clough Quartzite. The rocks are gray weathering schists containing quartz-muscovite-chlorite-biotite-garnet-graphite-leucoxene. Typical of the lower- to middle-garnet zone in the region these rocks are fine grained and virtually slates. Bedding is difficult to discern in this unit. The prominent schistosity is folded by a local map-scale dome-stage fold (figure 2).

Outcrops approximately halfway up the pasture are metamorphosed conglomerates of the middle member of the Clough Quartzite. Here the conglomerates include a few schistose clasts as well as an impure, somewhat schistose matrix.

Continue to traverse towards the woods at the northwest corner of the lower pasture. Cross the barbed-wire fence carefully and continue northwards along the logging road. Abundant outcrops, especially on the west side of the logging road, are clean, quartz-pebble conglomerates with quartzite matrix. The conspicuous pebbles are uniformly vein quartz. This lithology

is typical of the Clough Quartzite as mapped by most workers and is representative of the middle member of the Clough Quartzite throughout the Bernardston-Chesterfield area.

Proceed northward along the logging road to where the road takes a nearly right-angle bend to the west. Marbles of the Fitch Formation are exposed in several small pits dug during mining of a small magnetite bed in the eighteenth and nineteenth centuries (figure 4). Most of these pits are under the hemlocks to the north of the road. Due to the limited amount of conodont-bearing material and the unfortunately rapid disappearance of these important outcrops please **DO NOT HAMMER OUTCROPS** at this locale. There is abundant loose material for collection. The Fitch here is chiefly white to light-gray calcite marble containing scattered crinoid ossicles and minor (typically 1-3 percent modally) quartz, epidote, and pyrite. The prominent bed of magnetite-chlorite-quartz-garnet granulite is present in several pits as much as 0.8 m thick. The fossiliferous section of the stratigraphically highest Clough is interpreted to be a metamorphosed marine deposit which contained stenohaline, marine invertebrates (Boucot and others, 1958). Although this stratigraphically highest part of the Clough has been completely removed from the outcrops, its position was clearly documented by geologists in the ninteenth century. The map pattern of the Fitch, both locally and regionally, suggests that the Fitch unconformably overlies the Clough. The Fitch is unconformably(?) overlain by gray schist and quartzite of the Littleton Formation. One small, ground outcrop of Littleton Formation is shown on the map in figure 4. The outcrop is unimpressive but important as the best constraint on the stratigraphic top of the Fitch at this locale.

Within the marbles lensoid bedforms, relict crossbedding(?), abrupt variations in grain size, and a wide range in the size of bioclasts indicate deposition in a near-shore, shallow-water, variable, but generally high-energy, marine environment. Megafossils include crinoid ossicles (as large as 3 cm in diameter), columnals (as much as 5 cm long), and large recrystallized coralline fragments. Taken together, these data indicate that the Fitch near Bernardston is chiefly a carbonate shoestring sand that formed as a channel filling, bar, or beach. The overall transition from Clough to Fitch to Littleton represents a grossly deepening upward sequence.

The southerly dip of the units at this location, coupled with the clear-cut stratigraphic distinctions and paleontological age constraints, presents straightforward evidence that the rocks in the region have been structurally inverted. This inverted belt of rocks continues for tens of kilometers along strike to the north and makes up the inverted limb of the Bernardston nappe. The Bernardston nappe is the structurally lowest of the nappe-stage Acadian folds at this latitude. Although outcrop-scale structural features unequivocally associated with the nappe-stage are rare, a small, south plunging nappe-stage fold in bedding can be seen on the north facing outcrop near the sample locality labeled 891-U1 in Figure 4.

**Please retrace your path back to your vehicles** being careful to **leave all fences and gates as you found them.**

Continue northward on Route 5.

1.7 Turn right onto Burke Flat Road.

1.9 Bridge over Interstate-91.

2.0 Intersection with Bald Mountain Road, bear right.

2.5 Outcrops of Clough Quartzite on the overturned limb of the Bernardston nappe behind the small red house on the left.

3.3 Junction Route 10, turn left (Route 10 north).

4.7 Historic marker - Lt. Ebenezer Sheldon's Fort.

4.9 Turn left onto Martindale Road.

5.5 Pass Shedd Road on right (road to Snow Hill staurolite locality), outcrops on SE corner of intersection are metamorphosed rhyolites of the Upper Member of the Ammonoosuc Volcanics. Continue straight on Martindale Road.

6.0 Pavement ends, continue northward.

6.6 End of maintained road, park vehicles and walk northwards on undeveloped extension of Martindale Road into the woods. At fork in road turn left to outcrop in woods overlooking road.

## STOP 2. Upper Member of Clough Quartzite (Metamorphosed Volcanics)

(approximately 20 minutes)

This short stop is to see the distinctive feldspathic biotite granulites that occur at several localities between the middle member of the Clough Quartzite and the Littleton Formation. The hope is to alert other workers in the region to the possibility of finding volcanic rocks within the Silurian-Devonian section. These rocks were originally mapped by Balk (1956) as a "metatuff" member of the Bernardston Formation. I agree with Balk that the fine grain size, compositional layering, and mineralogy suggests these are metamorphosed felsic volcanic rocks. Although I have informally mapped this unit as an upper member of the Clough Quartzite, its stratigraphic position with relation to the Fitch Formation has not been established. Two aspects to note at this stop are the exposed contact with the Littleton Formation on the north and west sides of the small knob of outcrop and the unusual haloes of biotite-depleted matrix surrounding garnets.

**Return to vehicles.** Turn vehicles around and return south on Martindale Road.

7.2 Pavement begins.

8.3 T-intersection, turn left onto Route 10.

9.0 Turn right onto Deacon Parker Road.

9.1 Turn right into upper driveway of Valley Masonry and Construction Co. Bear left and proceed to southern parking lot.

9.2 Park in southern parking lot. Outcrops on east side of lot.

## STOP 3. Ammonoosuc Volcanics in Core of West Northfield Anticline

(approximately 1 hour)

The abandoned quarry outcrop on the east side of this parking lot is the largest exposure of the Lower Member of the Ammonoosuc Volcanics in the Bernardston-Northfield area. The rocks are fine-grained greenstones and represent metamorphosed tholeiitic basalts. A few very thin (<1 cm) calcite layers may be marble beds similar to those reported in the Lower Member of the Ammonoosuc by Schumacher (1985, 1988). These rocks have a much "lower-grade" texture than those east of the Connecticut Valley border fault.

Walk around the southwest end of the outcrop and traverse eastward, up the hill to Deacon Parker Road.

Walk south along Deacon Parker Road approximately 55 meters. Cross the road and enter the woods to the east. Proceed to moderately large outcrop overlooking road.

This outcrop exposes part of the Upper Member of the Ammonoosuc Volcanics. The rocks are felsic gneisses containing orthoclase, plagioclase, quartz, muscovite, and a little biotite. The rocks typically weather to an orange-brown stain and are closely jointed in the region. These are interpreted as metamorphosed rhyolitic volcanics. Beautiful volcanic textures are preserved in the Upper Member. The southern part of this outcrop contains the nicest volcanic textures at this stop. The rocks resemble welded tuffs or bomboclastic breccias. In some locations, the bombs contain relict blue-quartz phenocrysts.

The distictive thin Middle Member of the Ammonoosuc Volcanics (containing coticule) is exposed in the woods to the south of the quarry parking lot. The long walk and small outcrops preclude a visit to it on this trip.

**Return to vehicles.** Turn vehicles around and return to road.

| | |
|---|---|
| 9.3 | Turn left onto Deacon Parker Road. |
| 9.4 | Stop sign, go right onto Route 10 north. |
| 9.7 | Turn right, enter Northfield-Mt. Hermon School, Mount Hermon Campus. |
| 9.8 | Mount Hermon School gates, followed by outcrops of Mount Hermon Hornblende Gabbro and Partridge Formation on left and right. |
| 10.0 | Mount Hermon Hornblende Gabbro outcrop on right. |
| 10.2 | Bear right. |
| 10.4 | Turn left towards Main Campus. |
| 10.5 | Pass tennis courts on left. |
| 10.6 | Turn right, pass Manchester Hall on right. |
| 10.7 | Athletic field and track on left, outcrops of Mount Hermon Hornblende Gabbro in bleachers. |
| 10.75 | Pull off road to right, park across from Memorial Chapel. |

## STOP 4. Mount Hermon Hornblende Gabbro (approximately 30 minutes)

The low outcrops in front of Memorial Chapel on Mount Hermon represent the type locality for the Mount Hermon Hornblende Gabbro. The coarse relict gabbroic texture of these rocks is a stark contrast to that of the foliated, fine-grained amphibolites in the Lower Member of the Ammonoosuc Volcanics and the Partridge Formation in the field-trip region. The outcrops in front of the chapel are largely massive, but the rock is well foliated near contacts with surrounding stratified units. Although Trask (1964) mapped these coarse amphibolites as metamorphosed volcanics within the Partridge Formation, their texture, grain size, and the one continuous mappable region comprising the Mount Hermon pluton in the southern part of the area (figure 2), have led me to map them as a group of newly recognized metamorphosed plutonic rocks (Elbert, 1984). Numerous coarse dikes and sills of nearly identical gabbroic rocks occur within the Bernardston-Chesterfield area. I have mapped rocks I consider to be dikes of Mount Hermon Hornblende Gabbro in all the stratified units in the area except the Fitch Formation and the Littleton Formation. Although I have found these intrusives within a few feet of the Littleton Formation, I have been unable to

trace them into it; neither have I been able to show the unconformity at the base of the Littleton Formation cutting one of the gabbros. In spite of the latter, I believe the gabbro is a Silurian intrusive. Further, the gabbros are all located near or within the Brennan Hill thrust sheet and have not been found either within the overturned limb of the Bernardston nappe or in the tectonic levels above the Brennan Hill thrust sheet (figure 2 of Robinson and others, this volume). They seem to be concentrated along the zone representing the transition from the thin Silurian section deposited on the western margin of the Merrimack trough to the thick Silurian section deposited to the east.

Preliminary major- and trace-element geochemical characterization of Mount Hermon Hornblende Gabbro from the Mount Hermon pluton as well as from dikes throughout the country rock indicates the rocks have the compositions of olivine-normative basalts.

The geochemistry, texture, and map distribution suggests that the Mount Hermon Hornblende Gabbro was a mantle-derived melt which intruded the region of highest crustal extension and flexure related to the transition region of the Silurian basin. Similar rocks occur elsewhere in the Bronson Hill anticlinorium-Merrimack synclinorium (notably the Skitchewaug Mountain area) and their mapping may provide more clues to the early tectonics of the region. These gabbroic rocks may also have been an important source of heat in the basin and may have exerted a heretofore overlooked control on the metamorphic development of the region.

| | |
|---|---|
| | **Return to vehicles.** Turn around and drive back past the athletic field. |
| 10.9 | Turn right at intersection |
| 11.0 | Pass Schauffler Memorial Library on left. |
| 11.2 | Intersection, turn left – leave main Mount Hermon campus, continue past North Farmhouse on left. |
| 11.4 | Mt. Hermon Hornblende Gabbro outcrops on right in pasture. |
| 11.6 | Stop sign at T-intersection, turn left. |
| 11.8 | Northfield town line. |
| 11.9 | Stop sign at intersection with Route 10. Turn right. |
| | (Note: Outcrops just a few hundred feet to the southwest, left, on Route 10 are rusty weathering pelitic schists and amphibolites of the Partridge Formation. Those on the northern side of the road include a Mesozoic shear zone.) |
| 12.5 | Bennett Meadow Bridge over Connecticut River, skyline view of Notch Mountain, site of stratigraphic studies of the Ammonoosuc Volcanics (Schumacher, 1981, 1983, 1988) in the mantle of the Warwick dome, in the Mt. Grace quadrangle. |
| 13.0 | Cross small bridge, outcrop of Lower Jurassic, conglomerate member of the Turners Falls Sandstone in the small, Northfield basin located down the embankment on the northeast side of the bridge. |
| 13.2 | Flashing light and stop sign at junction of Route 63, turn left onto routes 10 and 63 north. |
| 13.9 | Flashing light in center of Northfield, IGA supermarket on left. |

| | |
|---|---|
| 14.5 | Pass Mill Street on left, reputed home of Newell Trask during field mapping in Bernardston-Chesterfield area in the early 1960's. Continue north on routes 10 and 63. |
| 14.6 | Pass low outcrop of Lower Jurassic, conglomerate member of the Turners Falls Sandstone in elevated driveway on left. |
| 15.1 | Traffic light, Northfield-Mt. Hermon School, Northfield Campus on right. |
| 15.6 | Leaving Pioneer Valley... |
| 15.7 | Turn left on Route 63 north toward Hinsdale, N.H. |
| 16.3 | Entering Winchester, N.H.<br>Begin Dartmouth College Road:<br>*"Over this Route Eleazer Wheelock passed to found Dartmouth College, 1770."* |
| 17.2 | Pass "Bee Supplies" on left. |
| 17.6 | Under RR bridge, clearance 13'11". |
| 18.6 | Under RR bridge, clearance 12'. |
| 18.7 | Outcrop on right of rusty-weathering Partridge Formation. |
| 18.8 | Pull off road on left side onto wide gravel shoulder. Park and carefully cross Route 63. Walk back (southward) along the railroad bed to the large outcrop. |

**<u>STOP 5. Partridge Formation at Dole Junction</u>** (approximately 30 minutes)

The Partridge Formation in the core of the Bernardston nappe includes rusty-weathering pelitic schists, amphibolites, and felsic gneisses. The amphibolites represent metamorphosed basaltic volcanics while the felsic gneisses are metamorphosed dacites (Hollocher, 1985). In light of the newly recognized presence of intrusive mafic rocks in the region (specifically the Mt. Hermon Hornblende Gabbro), the presence of felsic rocks within the Partridge Formation can be especially helpful in distinguishing the Partridge Formation from the rusty-weathering parts of the Silurian Rangeley Formation which do not contain metamorphosed volcanics. In addition to staurolite-garnet-biotite schists, this outcrop contains a meter-thick felsic-volcanic bed.

| | |
|---|---|
| | Return to vehicles and continue north on Route 63. |
| 19.1 | Local color to your left, hold on to your license plates... |
| 19.8 | Pass outcrops of Pauchaug Gneiss in core of Vernon dome on right. |
| 20.4 | Turn right onto Tower Hill Road (the turn is somewhat obscured when approached from the south). |
| 20.6 | Former site of railroad bridge. |
| 20.8 | Pass Depot Street on left, outcrops behind house northwest of intersection are parautochthonous Ammonoosuc Volcanics on the Vernon dome, small |

outcrop at northeast corner of intersection is Clough Quartzite on the dome. Continue up Tower Hill Road.

| | |
|---|---|
| 20.9 | Pass Old Northfield Road on right, continue up Tower Hill Road. |
| 20.95 | Cross trace of Clough Quartzite on overturned limb of Bernardston nappe, continue up Tower Hill Road. |
| 21.05 | Outcrops of Partridge Formation in the core of the Bernardston nappe in brook to left of road. |
| 21.6 | Turn vehicles around in dirt road on left, drive back down Tower Hill Road. |
| 21.7 | Pull off road on right and park. |
| | Walk northwestward (follow flagging) to the western summit of Biscuit Hill. |

**STOP 6. Biscuit Hill** (approximately 2 hours)

Biscuit Hill presents an exceptional opportunity to see the Monadnock-western Maine sequence stratigraphy. A significant portion of my 1985 field season was spent mapping this hillside and the surrounding region at 1:1000 on a 100-foot grid base map prepared by tape and Brunton survey. This trip's traverse has been designed to proceed from the top to the bottom of the hill, thereby working ***structurally downwards*** and ***stratigraphically upwards***. **Please be careful, the hillside is deceptively steep and there is a lot of loose rock to slip on or to knock onto other participants. Please hammer with forethought. Important folds, graded beds, and rare rock types are sometimes hard to see and should not be destroyed. If you are not absolutely sure you're sampling benignly ... ask for help!** There is quite a bit of collectable material already loose.

A description of the important stratigraphic, structural, and petrologic aspects of this stop is given in the body of this chapter. The following is meant to be a specific reminder of what to look for on the hill.

Most of the details of the Biscuit Hill stratigraphic sequence can be seen on this hillside traverse. They include the following: Interbedded gray-weathering schist with schist-matrix conglomerates and rare calc-silicate pods of the Rangeley Formation. A few thin interbedded quartzites mark the stratigraphic top of the Rangeley Formation and the gradation into the cyclically interbedded quartzites and gray schists of the Perry Mountain Formation. Although the contact between the Rangeley and Perry Mountain formations is gradational, exposure is sufficient in this region to map it consistently using the criteria that all conglomerate, calc-silicate pods, and epidote-bearing quartzites are mapped within the Rangeley Formation, while the base of the Perry Mountain Formation is marked by the first, continuous quartzite bed stratigraphically above schists and quartzites containing these distinctive Rangeley lithologies. Graded beds near the Perry Mountain-Rangeley contact confirm the topping direction. The Perry Mountain contains biotite-rich, massive gray schist with beds and boudins of fine-grained coticule (garnet granulite) and magnetite-grunerite-garnet-apatite-quartz-graphite iron formation. Stratigraphically higher rocks at Biscuit Hill are well bedded, rusty-weathering, sulfidic, graphitic calc-silicate granulites and interbedded sulfidic Mg-biotite schists of the Francestown Formation. These are stratigraphically overlain by well bedded, clean actinolite-garnet-calcite calc-silicate gneisses, garnet granulites and interbedded purplish biotite granulites of the Warner Formation. A pair of large sills of metamorphosed gabbroic rock occur on the side of the hill. Although these are substantially altered and largely composed of biotite rather than hornblende (one also contains coarse garnet porphyroblasts), I have mapped them as sills of Mount Hermon Hornblende Gabbro, the only gabbroic intrusive known in the vicinity.

Backfold- and dome-stage deformations are recorded in outcrop-scale folds, rock fabrics, and map pattern. These two phases have virtually parallel axial surfaces as well as fold axes that plunge moderately towards the southeast, parallel to the mineral and pebble lineations, and are difficult to separate. However, some minor folds can be shown to have axial planes parallel to the prominent foliation, and are backfold stage, while others fold that same foliation, and are dome stage. Structural analysis is impeded by the presence of abundant magnetite-bearing iron formation in both outcrop and float. These two phases were preceded by early isoclinal folding and then thrusting which first produced the overall inversion of the Biscuit Hill sequence of rocks and then juxtaposition of the uppermost Silurian Warner Formation directly against the Middle-Ordovician Partridge Formation in the core of the Bernardston nappe just south of the base of the hill. Field trip participants may want to argue about what I feel to be the remnants of a fabric which may have been associated with the early isoclinal folding and can be seen in the beautifully graded beds in the Perry Mountain Formation.

Walk out (south) along logging road at base of Biscuit Hill, then ~0.25 miles uo (east) Tower Hill Road. Follow flagging southward a few hundred feet into the woods to the exposure of the Brennan Hill thrust.

The Brennan Hill thrust itself is not exposed on Biscuit hill. It is exposed here, a short distance to the south placing Upper Silurian Warner Formation against Middle Ordovician Partridge Formation. One of the most interesting features of this fault is that it must be identified by stratigraphic mapping. The fault predates significant metamorphic recrystallization and a preserved fabric has not been identified.

**Return to Tower Hill Road and walk up (east) to vehicles.**

Continue driving down Tower Hill Road.

| | |
|---|---|
| 22.0 | Pass logging-road exit from Biscuit Hill. |
| 22.3 | Turn right onto Depot Street. |
| 22.4 | Outcrop of Ammonoosuc Volcanics on left. |
| 22.6 | Drive under power line which passes over Cannon Hill. |
| 22.7 | Across defunct railroad crossing, continue down Depot Street. |
| 22.8 | Cross bridge over Ashuelot River. |
| 23.0 | Junction routes 119 and 63, cross Route 119 and follow Route 63 north towards Chesterfield. |
| 23.1-23.8 | Outcrops on left and right of Ammonoosuc Volcanics on Vernon dome. |
| 26.9 | Pass logging-road entrance to Pisgah State Park on right. |
| 27.0 | Large roadcut of Partridge Formation on right. |
| 27.1 | Trace of Brennan Hill thrust location on road. |
| 27.2 | Outcrop of Rangeley Formation in Brennan Hill thrust sheet on right. |
| 27.8 | Chesterfield Town Line. |
| 28.7 | Pass Crowningshield Road on left, continue north on Route 63. |
| 28.9 | Pass North Hinsdale Road on left, continue north on Route 63. |

| | |
|---|---|
| 29.3 | Turn left onto dirt road, intersection at 229 meters elevation on Winchester, N.H. topographic quadrangle map. |
| 29.8 | Stop sign, turn left onto Castle Road. |
| 30.2 | Join Gulf Road (unmarked), continue straight. |
| 30.5 | Bear right at fork to stay on Gulf Road. |
| 31.3 | Pull off on right to park. It is difficult to park more than one vehicle here, but at the time of this writing construction had just begun on several houses along this road and there may be access roads by the time of the field trip. |

Walk into woods to the north and follow flagging to outcrops along south-southeast end of hill 345 (meters) which is southeast of Sargent Hill (Newfane, VT-NH, $7\frac{1}{2}$x 15 minute quadrangle), and in the northeasternmost corner of the Brattleboro, VT-NH $7\frac{1}{2}$x 15 minute quadrangle.

## STOP 7. Hill 345, South of Sargent Hill (approximately $1\frac{1}{2}$hours)

This final stop is a traverse from the core of the Bernardston nappe structurally upwards across the Brennan Hill thrust into a right-side-up sequence of eastern stratigraphy within the Brennan Hill thrust sheet (figure 8). The traverse starts with an outcrop of metamorphosed gabbro, mapped as Mt. Hermon Hornblende Gabbro. This gabbro strongly resembles the type outcrops seen at Stop 4, and highlights the regional extent and importance of these intrusives. The first stratified rocks on this traverse are rusty schists of the Partridge Formation. This belt of schist contains metamorphosed felsic volcanics along strike to the north and south, and does not contain any of the rock types (calc-silicate pods, quartzites, or conglomerates) indicative of the Rangeley Formation.

Traversing eastwards along the hillside the first contact is with gray- and rusty-weathering staurolite schists with interbedded conglomerates and quartzites of the Rangeley Formation. This contact is the Brennan Hill thrust. The path of the traverse drops to an elevation of approximately 240 m and continues 9 meters across strike to the contacts with the Perry Mountain and Francestown formations. Approximately 7.5 meters of Francestown Formation is then overlain by 12 meters of Warner Formation. The traverse continues across a three-meter covered interval to outcrops exposed at 236 m elevation. The section in the Brennan Hill thrust sheet is then repeated by a small imbricate splay. Here the rocks are staurolite schists with interbedded quartzites and conglomerates, of the Rangeley Formation, overlain by magnetite-apatite iron formation marking the Perry Mountain Formation, Francestown and Warner formations.

This stop shows both a small imbricate of the Brennan Hill thrust and a region where the thrust has emplaced right-side-up Monadnock sequence against the core of the Bernardston nappe. Several of these small splays, some involving Partridge Formation, have been identified in the Bernardston-Chesterfield area. Although this particular imbricate was followed for over three-quarters of a mile, detailed mapping of the splays is not routinely possible due to poor exposure and the preponderance of pegmatite and metamorphosed gabbro dikes and sills. These intrusions obscure contacts of stratified rocks and cut out an indeterminable amount of the stratigraphy. This is critical in this region where entire thrust imbrications can be demonstrated to be no more than thirty feet thick and contain stratigraphic units just a few feet thick.

The right-side-up belt of eastern sequence rocks at this stop are on the lower limb of an isoclinal syncline called the Hinsdale syncline (Elbert, 1986). The inverted section at Biscuit Hill is on the eastern, upper limb of the Hinsdale syncline. The axial surface of this early fold is truncated by the Brennan Hill thrust (figure 6), clearly identifying the fold as pre-thrusting. As can be seen in the northern cross section of the area (figure 9) the Hinsdale syncline is interpreted to be the same age as but structurally lower than the Monadnock syncline (Thompson, 1985, and this volume).

**Return to vehicles** and continue west on Gulf Road.

31.4 Hard left turn (140°).

31.9 Turn left onto Bradley Road (unmarked).

32.0 Road comes in from right, continue straight on Bradley Road.

32.7 Stop sign. Go right, rejoining Gulf Road headed northeast.

33.1 Bear right at fork.

33.4 Cross Orchard Road; keep going northeast.

33.7 Pass Noyes Robertson Coolidge Cemetery on left.

34.0 Stop sign, intersection with Stage Road, turn right.

34.1 Stop sign at T-intersection with Route 63, turn left (north) towards Spofford.

34.7 Center of Chesterfield – birthplace of Harlan Fiske Stone (1872-1946), Chief Justice of the United States Supreme Court, 1941-1946. Continue north on Route 63.

35.8 Stop sign at T-intersection of Route 63 with Route 9. Turn right (east) and follow Route 9 approximately 9.5 miles to Keene, NH.

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NEIGC FIELD TRIP A-6

OCTOBER 14, 1988

# DEGLACIATION OF THE CONNECTICUT VALLEY: VERNON, VERMONT, TO WESTMORELAND, NEW HAMPSHIRE

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## INTRODUCTION

The main purpose of this field trip is to study the glacial, late-glacial and postglacial history of the Connecticut Valley between Vernon and Westmoreland. There are several interrelated themes that will be addressed. They include the direction of ice movement, the mode of ice retreat, morphosequences, glacial Lake Hitchcock, post-Lake Hitchcock sediments and crustal rebound. Field trip stops will be made on the following U.S. Geological Survey 7.5'x 15' quadrangles with a metric scale of 1:25,000: Brattleboro, Vt-NH; Newfane, Vt-NH, and Keene, NH-Vt.

Weathering and erosion by streams and continental ice sheets of complex metamorphic rocks intruded by granite have produced a rugged, hilly topography with local relief of 350 m (1,150 ft). The low elevation for most of the field trip area is supplied by the Connecticut River which is a lake, the surface of which is held at 66 m (217 ft) ASL by the Vernon dam. The local high elevation is 417 m (1,368 ft) ASL on Wantastiquet Mountain just east of the Connecticut River at Brattleboro. Stream drainage is controlled by the Connecticut River which flows south and has two major tributaries in this area, the West and Ashuelot Rivers.

The area probably has been covered by ice sheets several times but specific evidence of multiple glaciation in southeast Vermont and southwest New Hampshire is sparse. A loose sandy till overlying very compact till with stained joints can be observed at two localities near Keene. Very compact till with stained joints occurs at several sites in the Newfane and Keene metric quadrangles, including STOP 9 of this field trip. Multiple-till exposures probably representing two separate glaciations are known in northern, east-central, and southern New Hampshire (Koteff and Pessl, 1985), as well as in southern Quebec and southern New England.

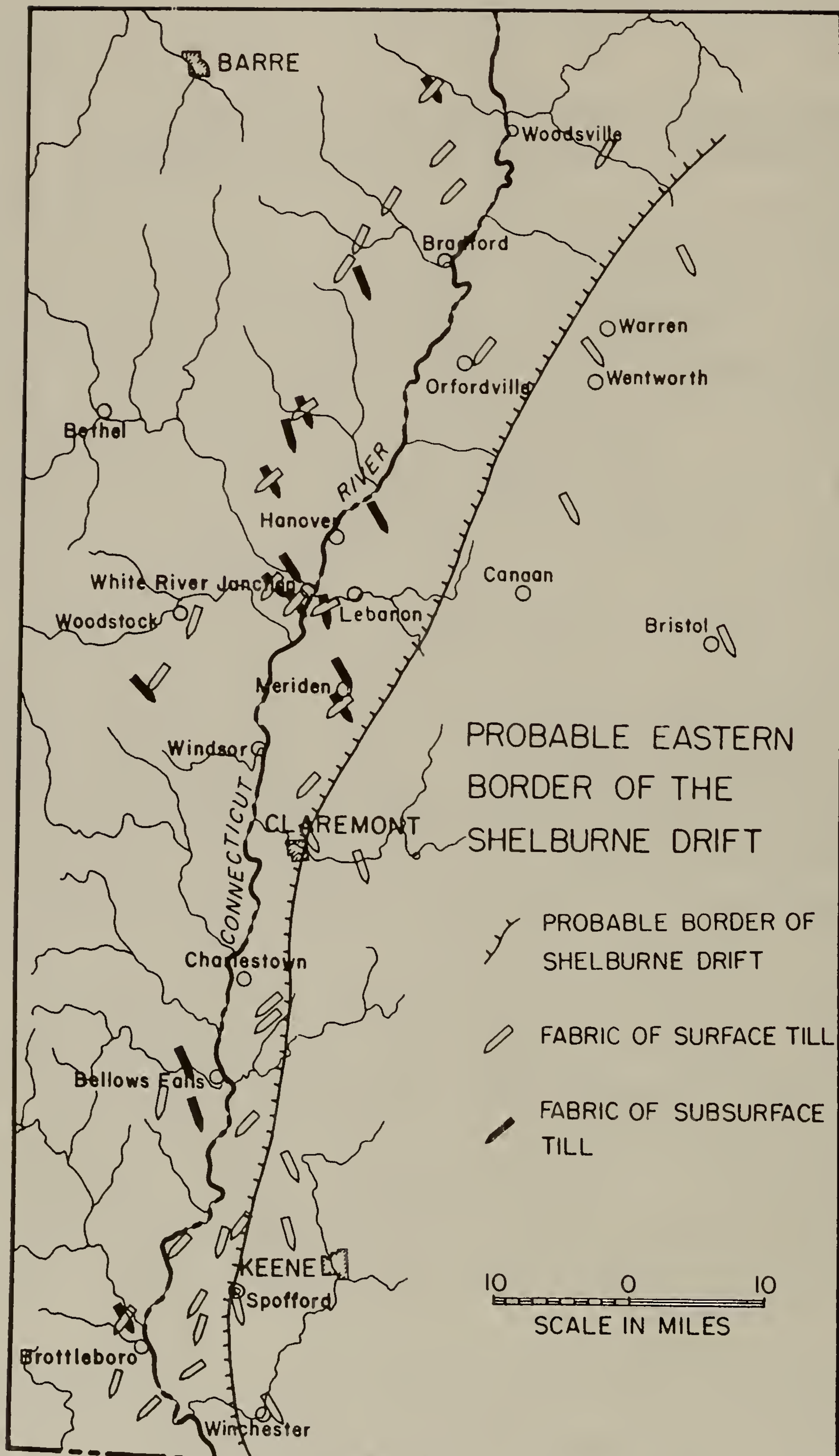


Figure 1. "Probable eastern border of the Shelburne drift". According to Stewart and MacClintock (1969, p.99), the Shelburne drift presumably was derived by ice moving from the northeast. Note that surface-till fabrics west of the border between Winchester and Charlestown indicate that the Shelburne drift came from older drift east of the border. That situation would be highly unlikely if not impossible.

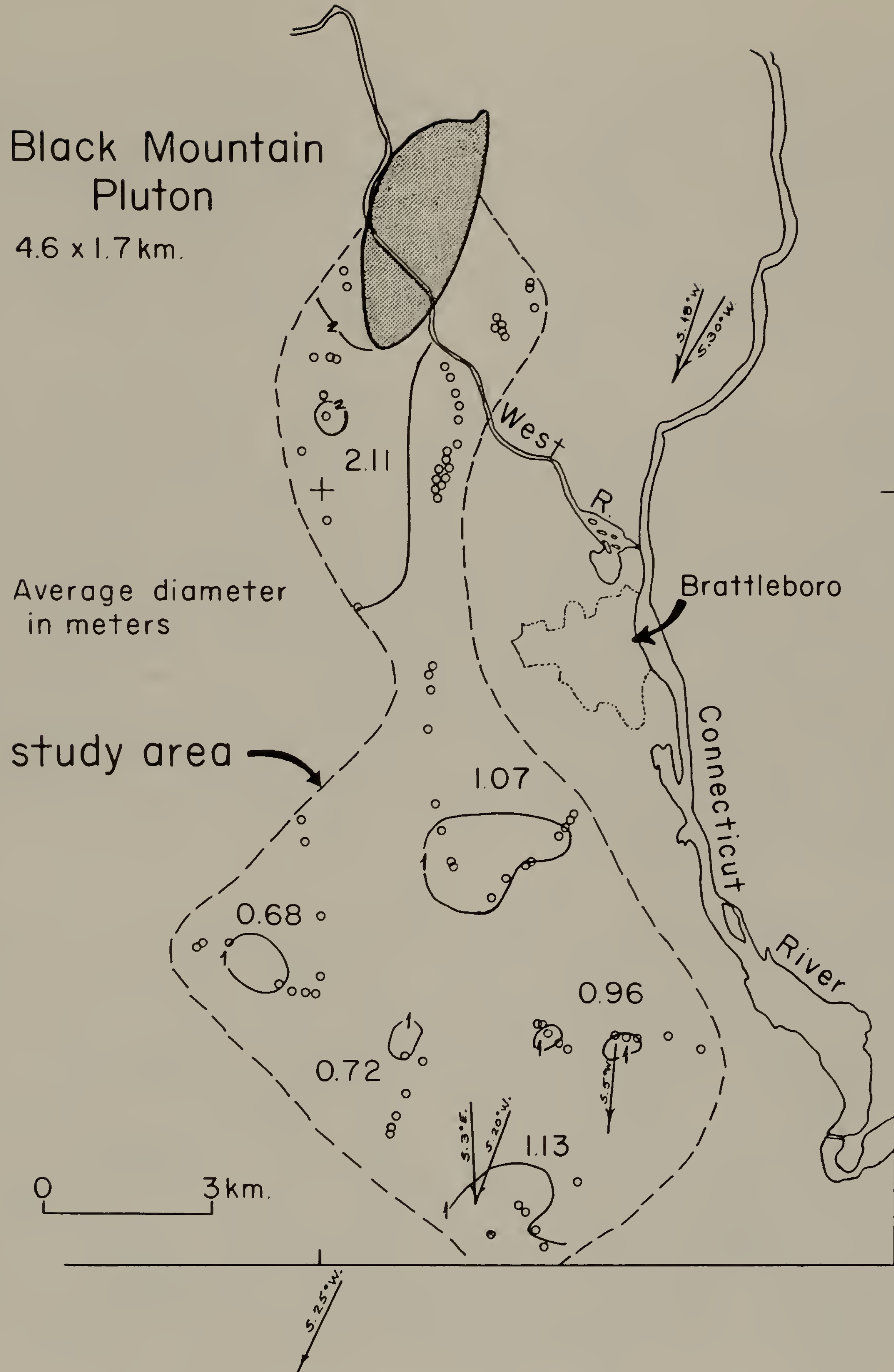


Figure 2. Distribution of glacial erratics of Black Mountain granite (open circles). Erratics were measured within a limited study area (dashed line). Numbers represent the average diameter of the group of erratics where the number is plotted. Arrows represent striations.

The margin of the last ice sheet retreated northward from Long Island at least by 19,000 years ago (Sirkin, 1982), and the Quebec Appalachians were deglaciated by 12,500 years ago (McDonald and Shilts, 1971). Therefore, the ice margin retreated through the field trip area between 19,000 and 12,500 years ago. Using linear interpolation and assuming a steady rate of ice margin retreat, we can guess that the ice margin retreated through the area about 15,000 years ago. Since retreat of the ice margin was not uniform, the 15,000 years is only a crude estimate.

Stewart (1961), and Stewart and MacClintock (1964, 1969, and 1970) recognized three separate drift (till) sheets in Vermont and westernmost New Hampshire. They named (1) a northwest-derived Bennington drift, (2) a northeast-derived Shelburne drift, and (3) a northwest-derived Burlington drift. All of the area of this field trip lies in the area of the so-called Shelburne drift (Fig. 1). The Shelburne drift was found not to exist at its type locality (Wagner, Morse and Howe, 1972) and 9 indicator fans mapped in the area of the Shelburne drift are oriented to the south-southeast and south (Larsen, 1987). According to Stewart and MacClintock, they should be oriented to the southwest of their source areas which they are not. We believe that the model of three drift sheets in Vermont and New Hampshire is untenable and that the surface till of New England resulted from the advance and retreat of one ice sheet, the Late Wisconsinan (Laurentide) ice sheet.

During retreat of the ice sheet in this area, the ice margin was accompanied by a northward-expanding glacial Lake Hitchcock (Lougee, 1939, 1957). Lake Hitchcock developed a stable outlet over a bedrock threshold at New Britain, Connecticut, and drainage down the present-day course of the Connecticut River was blocked by a large ice-contact delta at Rocky Hill, Connecticut (Stone and others, 1982). Recent work by Koteff and Larsen (1988) indicates that the former shoreline of Lake Hitchcock now rises toward about N21.5 W with a gradient of 0.90 m/km (4.74 ft/mi). The lake formed during ice retreat when the land was still depressed by the weight of the ice, and it extended 320 km from its spillway to West Burke, Vermont, before uplift due to the removal of the weight of the ice sheet commenced. Deltaic and lake-bottom deposits associated with Lake Hitchcock will be observed at several stops on this field trip.

## DIRECTION OF ICE MOVEMENT

The direction of movement of the former ice sheet in the Brattleboro area can be ascertained by a study of striations, roche moutonnée forms, and the distribution of granitic erratics from the Black Mountain pluton (Fig. 2). Striations, including some not shown on Figure 2, occur in two modal groups. One is S5E to S5W and the other is S20W to S35W. Due south-trending


Figure 3. A portion of James W. Goldthwait's compilation of glacial striations in New England; P, Putney, Vt., (Flint, 1957, p. 60).

striations cut by a younger set trending S40W can be observed at a site 0.7 km (0.43 mi) S86E of the east end of the Route 9 bridge over the Connecticut River.

The diameter of granitic erratics was measured in a limited study area south of the Black Mountain pluton (dashed line, Fig. 2). The average diameter in meters of small groups of erratics is plotted on Figure 2. In the study area south and southwest of Brattleboro, groups of larger erratics with average diameters of 1.07 and 1.13 m lie S5E of the Black Mountain pluton. Groups of smaller erratics lie both east and west of the area of larger erratics. In the study area northwest of Brattleboro, a group of erratics with an average diameter of 2.11 m is located southwest of the pluton. The two largest erratics measured, 3.66 and 4.06 m, were found just west of the pluton.

The above data from striations and indicator clasts can interpreted to indicate two separate phases of glacial movement. The first phase was characterized by essentially due south movement during glacial maximum. The second phase occurred in late-glacial time when an active ice lobe retreated northward through the Brattleboro area, forming the southwest-trending striations and pushing large granitic erratics to the southwest from the Black Mountain pluton.

In Massachusetts, deglaciation of the Connecticut Valley was by an active lobe of ice that readvanced several times (Larsen and Hartshorn, 1982). The active lobe is also indicated by a radial pattern of striations stretching across the valley and the distribution of erratics of Jurassic-Triassic rocks transported both east and west of their source area in the Connecticut Valley. A compilation of striations in New England made by James Goldthwait (Fig. 3) shows southwest and west-southwest striations located west of the Connecticut River in both Massachusetts and Connecticut.

A basic question then is how far north was the Connecticut Valley ice margin an active, spreading lobe? We believe that the answer lies on the Goldthwait map near Putney, Vt (P, Fig. 3), at the site of a striation that trends about S30W. On the west side of the Connecticut Valley north of Putney there are no striations to indicate a radial pattern of movement by an active ice lobe. Instead, there is a consistent south-southeast trend of striations from east-central Vermont to southwest New Hampshire. The lack of a radial pattern of striations seems to indicate that deglaciation north of Putney was by a stagnant tongue of ice. The width of the stagnant zone probably was many kilometers wide and, upglacier from the stagnant ice, the active ice was sluggish at best, showing no sign of lobate flow in late-glacial time.

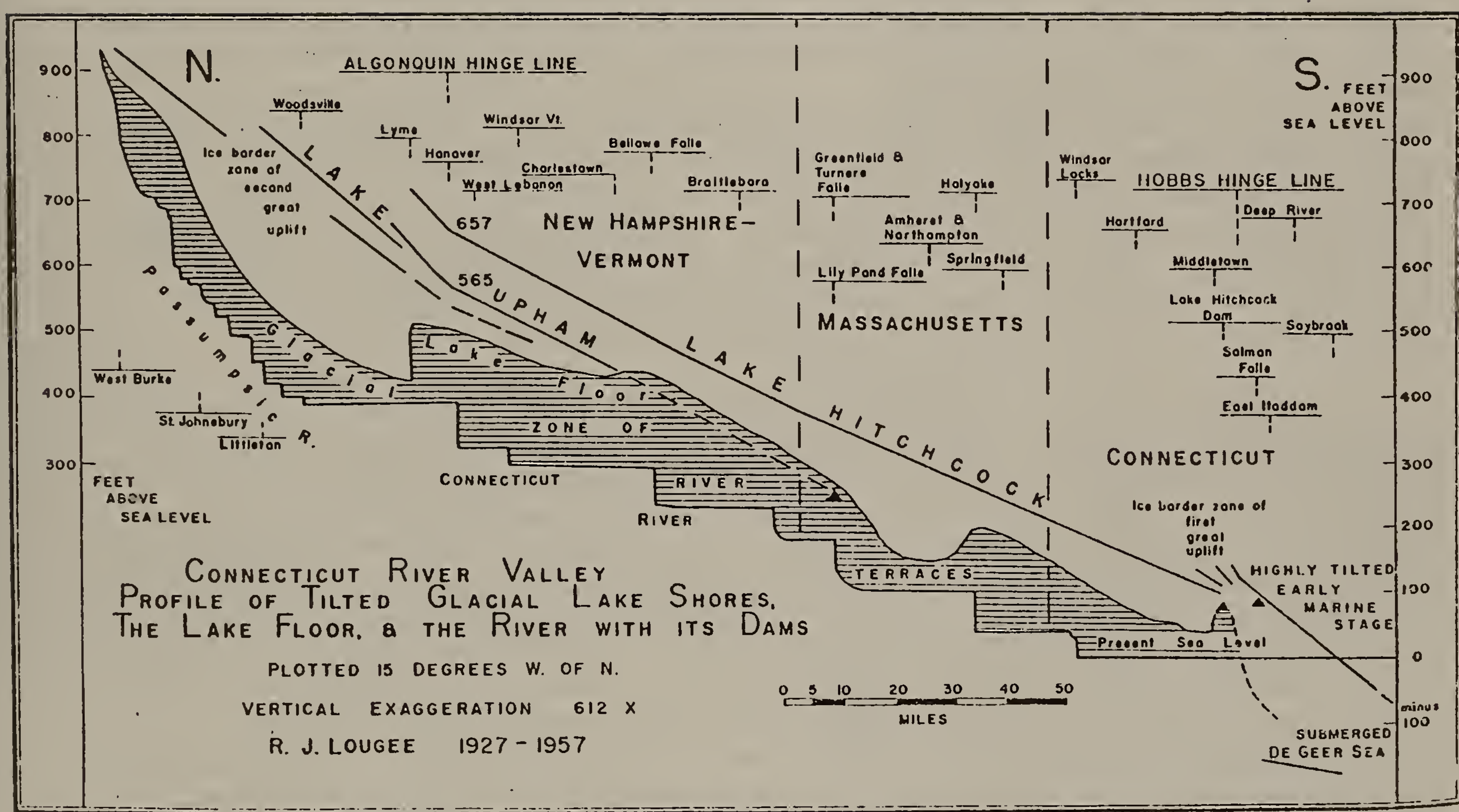


Figure 4. Lougee's (1957) projected profile of the levels of Lake Hitchcock and Lake Upham. Compare with Fig. 6. (From Dartmouth Alumni Magazine, November, 1957, p. 37)

## MORPHOSEQUENCES

A change in the regimen of the retreating ice sheet when its margin was near Putney, as suggested by the ice-directional features, can also be seen in the evidence from morphosequences.

A morphosequence consists of all the stratified deposits formed by one meltwater stream system and deposited in a given depositional basin defined by a particular ice margin and outlet from that basin (Larsen and Hartshorn, 1982; see Koteff, 1974, and Koteff and Pessl, 1981). A typical morphosequence is an ice-contact delta with coarse sediment at its proximal (ice-contact) side that grades downstream to fine sediment at a more distal (lacustrine) side. Retreat of the ice margin from the ice-contact slope marks the end of construction of a given morphosequence. When the ice margin again becomes stationary, or if there is a sudden outburst of meltwater (jökullaup), construction of a new morphosequence may start.

For the purposes of this field trip, we consider the main morphosequences near the axis of the valley that were built by meltwater flowing directly from the ice and were graded to a single lake, Lake Hitchcock, whose outlet was controlled by a stable threshold at New Britain, Connecticut. As the ice margin retreated northward up the Connecticut Valley near Brattleboro, at least 10 ice-contact deltas (morphosequences) were formed sequentially from south to north. An orderly northward retreat of the margin of the last ice sheet has been documented by many geologists throughout southern New England by the mapping of morphosequences (Koteff, 1974), so the presence of numerous morphosequences at Brattleboro is not unusual. What is unusual is the fact that near the axis of the Connecticut Valley north of Putney there are very few morphosequences built directly from the ice and there is no place that matches the number and close spacing of morphosequences as can be observed at Brattleboro.

It appears that the vitality or the activity of the Connecticut Valley lobe dropped considerably when the ice margin was located near Putney. Borns and Calkin (1977) have suggested that, when the Laurentide ice sheet thinned over the Appalachian Mountains of northwest Maine to the point where the ice surface strongly intersected the regional topography, then the ice sheet south of the mountains lost a regional supply of ice from the north and its regimen dropped sharply. Borns and Calkin further note that the large esker systems in Maine were formed in this zone of sluggish ice south of the mountains. That same zone extends southwest into Vermont and New Hampshire and includes the large Connecticut Valley esker that extends from Windsor, Vt, to Lyme, NH. Whether or not the rate of retreat of the ice margin in the Connecticut Valley can be related to the lowering of the ice sheet over the Green Mountains remains to be seen. It does appear that the ice sheet had a diminished ability to supply sediment to the ice margin when it retreated north of Putney.


Figure 5. Map of Lake Hitchcock and Lake Upham by Lougee (1957, from Dartmouth College Alumni Magazine, November issue, p. 27).

## HISTORICAL NOTE PERTAINING TO THE VERMONT-NEW HAMPSHIRE PORTION OF LAKE HITCHCOCK

(1) Richard J. Lougee's projected profiles of Lake Hitchcock and Lake Upham are oriented "15 degrees west of north" (Fig. 4) (Lougee, 1957). This is not far from its probable true value near N21.5W as shown by Koteff and Larsen (1985, 1988). Lougee thought that Lake Hitchcock drained when its northern end was located at the ice margin near Lyme and that Lake Upham, a lower and younger lake, extended north to Burke, Gilman and Littleton (Fig.5). Lougee extended the "Algonkian hinge line" across the Connecticut valley at Lyme. Lougee's "Lake Upham" north of the hinge line plus his "Lake Hitchcock" south of the hinge line essentially is the same as the Lake Hitchcock shown by Koteff and Larsen for Vermont and New Hampshire (Figs. 6 and 7). We see no evidence for hinge lines in the area of Lake Hitchcock and believe that Lougee came within a hinge line (or two) of unraveling the true story of Lake Hitchcock in 1957.

(2) Stewart and MacClintock (1969, 1970) on the Surficial Geologic Map of Vermont show a "Connecticut Valley Lake" extending along the entire Connecticut valley of Vermont from Massachusetts (at elevation 435 ft) to the International Border near Canaan (at elevation 1,101 ft). They assumed that maximum uplift was due north with a gradient of 4.15 ft/mi (Stewart and MacClintock, 1969). Such a lake as the "Connecticut Valley Lake" shown by Stewart and MacClintock on the Surficial Geologic Map of Vermont never existed.

(3) Larsen (1983) measured the elevation of 11 topset-foreset contacts between Brattleboro and Lisbon. It became apparent in 1983 that Lake Hitchcock extended to Burke, Gilman and Littleton, and was similar in those areas to the map of "Lake Upham" by Lougee (Fig. 5). Based on the 11 data points in Vermont and New Hampshire plus 4 others in Massachusetts, it was noted that the best-fit plane for the uplifted Lake Hitchcock shoreline rose toward N20W at 4.54 ft/mi (Larsen, 1983). That study has since been extended and now includes 28 data points on or near the Lake Hitchcock water plane, which is now thought to rise 4.74 ft/mi toward N21.5W (Koteff and Larsen, 1988).

## GLACIAL LAKE HITCHCOCK

As the ice margin retreated through the Brattleboro area it was accompanied by a northward-expanding Lake Hitchcock. For an up to date treatment of the origin and early history of Lake Hitchcock see Stone and others (1982) and Koteff and others (1987). By the time the ice margin was in the vicinity of the Holyoke Range in Massachusetts, the level of Lake Hitchcock had become stable because downcutting at the New Britain spillway had reached bedrock and ceased. Koteff and Larsen (1988) have established the location and orientation of the stable shoreline of Lake Hitchcock by measuring the elevation of the topset/fore-

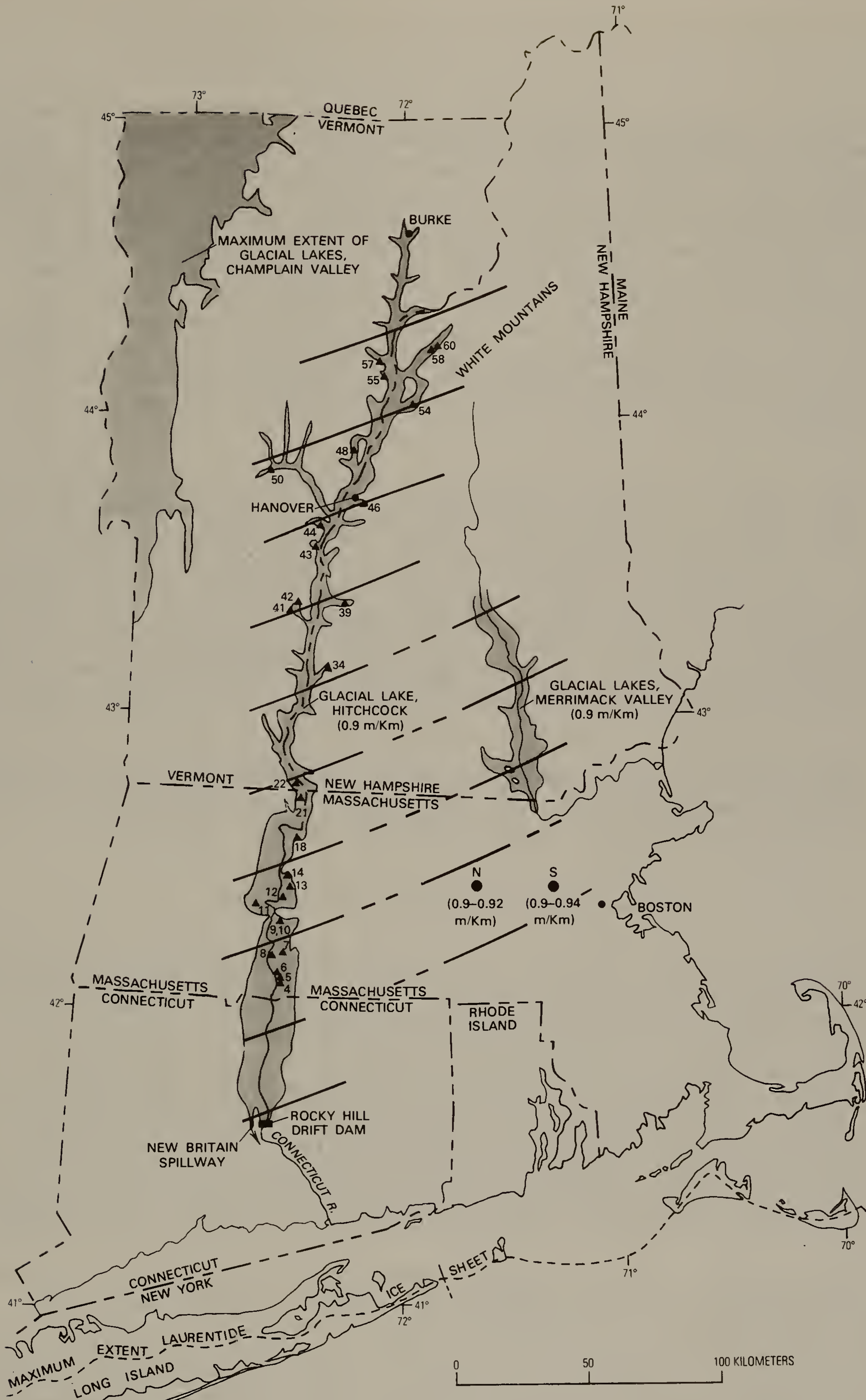


Figure 6. Generalized outline of glacial Lake Hitchcock and selected other glacial lake areas in western New England. (N) glacial Lake Nashua; (S) glacial Lake Sudbury; solid triangles denote location of altitude obtained from unmodified, ice-marginal, or meltwater-derived delta; uplift isobase interval is 25 meters. (Figure from Koteff and Larsen, 1988)

set contact in many deltas (Fig. 6). The highest 28 deltas are ice-marginal features, built consecutively from south to north, and define the former shoreline as a plane with a gradient of 0.90 m/km (4.74 ft/mi) rising toward N21.5W (Fig. 7). Because the former shoreline appears to be planar, as opposed to being curved, we believe that postglacial rebound did not commence in New England until the ice margin had retreated north of West Burke (Koteff and Larsen, 1988). If rebound had caused the spillway to rise while ice still occupied the northern part of the Lake Hitchcock basin the youngest deltas would have formed in a rising lake. That would have produced a concave-up profile instead of the linear projected profile that we see in Figure 7.

Only one of the ice-contact deltas seen on this field trip (Loc. 22, Figs. 6 and 7), our STOP 1, has a topset-foreset contact that falls on "the water plane" of Koteff and Larsen (1988). The topset-foreset contacts at Localities 23 through 30 (except Locality 26), which includes ice-contact deltas built directly into Lake Hitchcock, plus another T-F-C which has just been measured at STOP 7, fall well below the upper line of Figure 7. This is a problem that has not yet been resolved, but this field trip will present an opportunity for all to offer their suggestions.

The sediments of Lake Hitchcock that will be seen on this field trip consist of a great variety of ice-contact, deltaic and lake-bottom deposits and their associated features.

## POST-LAKE HITCHCOCK SEDIMENTS

At its maximum extent, an arm of Lake Hitchcock extended up the valley of the Second Branch of the White River and into Williamstown Gulf in central Vermont. Gravel bars located below the level of Lake Hitchcock in the valley bottom near East Brookfield appear to have been formed by the outlet stream from glacial Lake Winooski after Lake Hitchcock drained (Larsen, 1984, 1987). This indicates that the dam for Lake Hitchcock at Rocky Hill had been breached while glacial ice still blocked the northwest-draining Winooski River in Vermont. The draining of Lake Hitchcock has to predate any shell-bearing Champlain Sea deposits and could have occurred around or before 13,000 BP.

Draining of Lake Hitchcock resulted in the erosion of lake-bottom sediments by the Connecticut River, which left wide stream terraces as it cut down below 66 m ASL. A typical stream-terrace deposit consists of 5 to 6 m of yellowish-brown fine to very fine sand over pebble gravel up to a meter thick. These sediments rest disconformably upon lake-bottom deposits. Wind activity was great shortly after Lake Hitchcock drained, as shown by extensive sand-dune deposits that lie southeast of major stream terraces at STOP 8 and STOP 10.

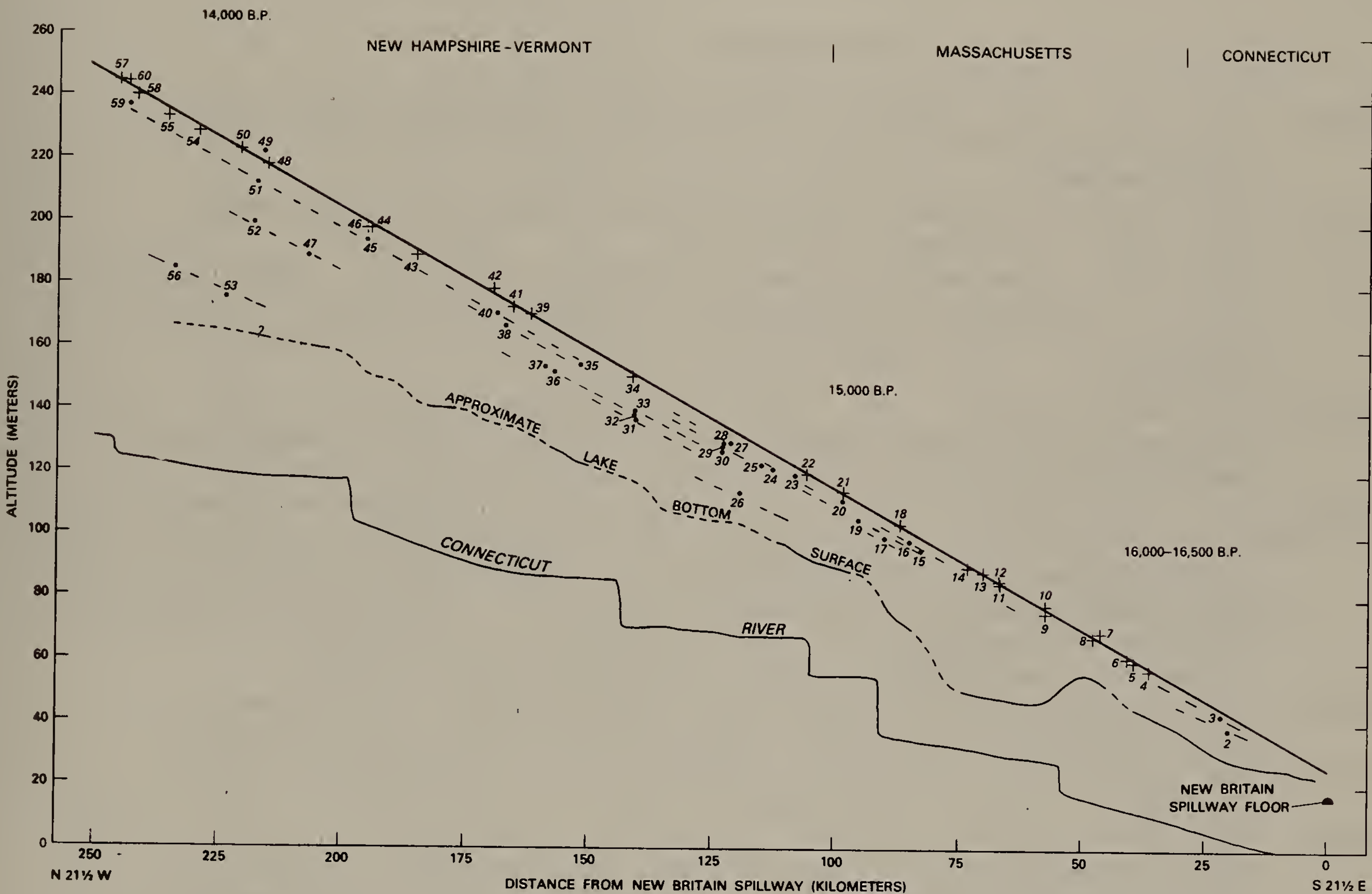


Figure 7. Ordinary least squares regression profile based on altitudes of topset/foreset contacts of 28 unmodified, ice-marginal, or meltwater-derived deltas (+) in glacial Lake Hitchcock. ( . ) other altitudinal data. Dashed profiles are diagrammatic only. Lake-bottom profile estimated from previous publications and topographic maps. (Figure from Koteff and Larsen, 1988).

## ACKNOWLEDGMENTS

Mapping by Larsen in the Newfane and Keene metric quadrangles was supported by Eugene L. Boudette, New Hampshire State Geologist. The work of Koteff and Larsen on the former level of Lake Hitchcock was supported by the U.S. Geological Survey. We thank Maureen Larsen for proofreading the paper.

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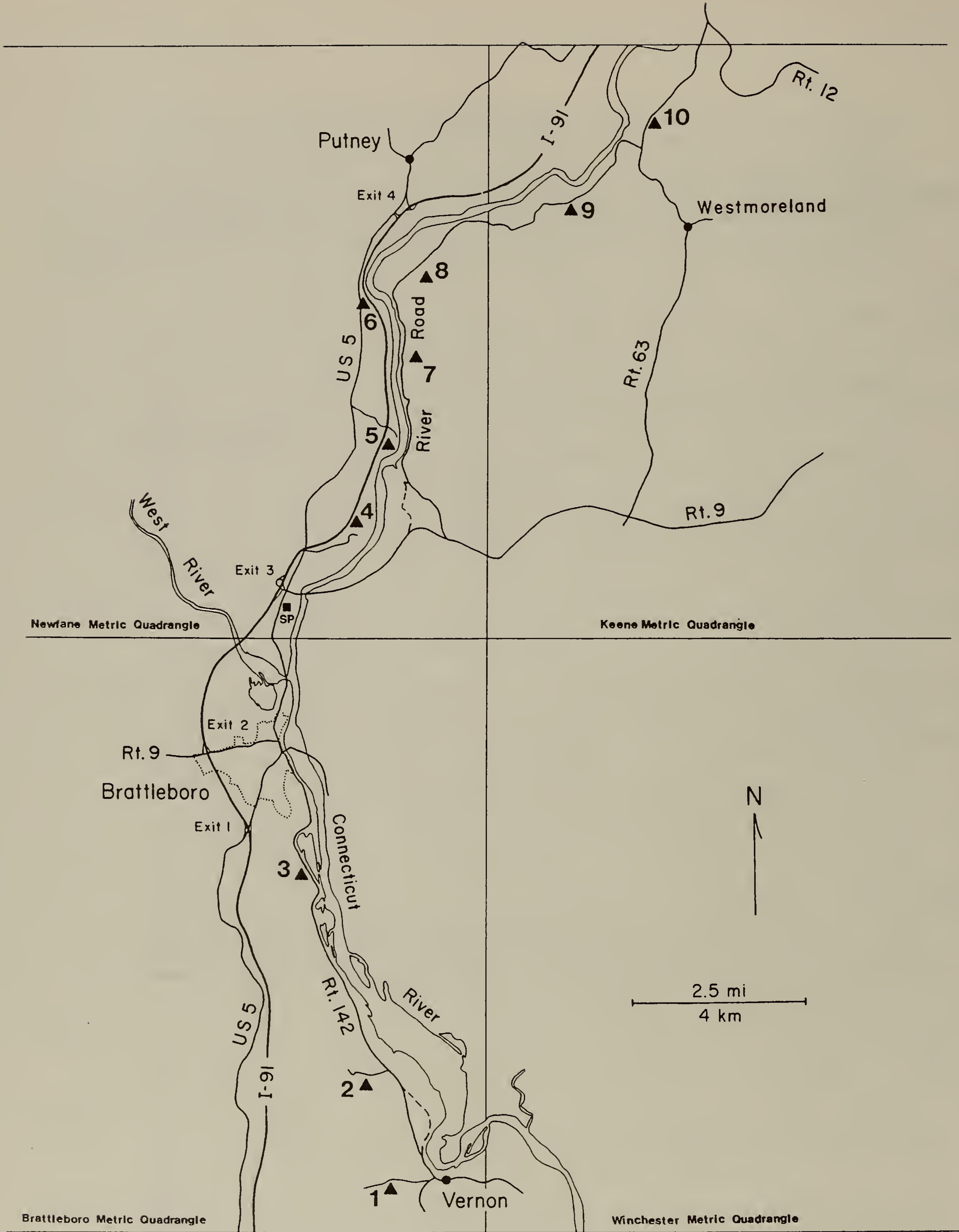


Figure 8. Location of field trip stops.

## ROAD LOG

START AT PARKING LOT OF MCDONALD'S RESTAURANT 0.25 MI SOUTH OF THE JUNCTION OF ROUTES 9 AND U.S.5 (NEXT TO EXIT 3 OF I-91) AT THE NORTH END OF BRATTLEBORO.

Mileage

0.0 Leave McDonald's Restaurant, turn left (south) on U.S.5, proceed to the south end of downtown Brattleboro

2.3 Turn left and immediate right on Route 142, proceed south along west side of Connecticut River

4.5 Cersosimo Lumber Yard, site of STOP 3, on right

9.4 Turn right (west) on Pond Road under railroad

9.45 Turn right on West Road, proceed west on stream-terrace deposits cut in Lake Hitchcock bottom sediments

10.00 Pit entrance on left. Reverse direction with caution and park off pavement

STOP 1. GASSETT PIT (Town of Vernon, Vermont): The pit is located in a linear ridge that trends N22W and rises over 120 m ASL (above sea level) on the Brattleboro, Vt-NH, 7.5' x 15' metric quadrangle (1:25,000). The southernmost face is over 280 m south of West Road and has in its upper part 2 to 3 m of pebble-cobble gravel with a matrix of coarse sand and granules. Low-angle crossbeds indicate transport direction to the southwest. About 60 m north of the south face deltaic foresets dip 22 degrees toward S75W. About 15 m further north beds of pebbly coarse sand, fine sand and silt display abrupt change in grain size, cut-and-fill structure, and opposing directions of transport to the northeast and southwest. In July, 1982, collapsed beds of lacustrine fine to medium sand were exposed 75 m north of the latter site. The general aspect of this landform is that of an ice-contact delta fed by meltwater streams issuing along an ice margin that trended north-northwest on the east side of the deposit. In 1982, the elevation of the topset-foreset contact in a former position of the south face was measured to be 119.5 m ASL (Koteff and Larsen, 1988).

Proceed east on West Road, turn left (east) on Pond Road

10.9 Turn left (north) on Route 142 just east of railroad underpass

12.7 Turn left (west) on Tyler Hill Road

12.95 Pit entrance on left

STOP 2. EVANS PIT: Town of Vernon, Vermont; the pit is located on Tyler Hill Road 2.78 km (1.72 mi) N33W of railroad underpass in Vernon (Brattleboro, Vt-NH, metric quadrangle).

The pit is in a ridge that rises toward S33W. About 305 m (1,000 ft) southwest of Tyler Hill Road and at an elevation of about 120 m ASL the ridge joins an irregular plain that slopes gently to the southwest. A small depression is located just southeast of the point where the ridge joins the plain. The southwest face has 1 m of pebble gravel with openwork structure over 1 m of fine to medium sand below which is a thick unit of poorly sorted pebble-cobble gravel with subangular to subrounded boulders. On the east flank of the ridge are younger fine-grained lake sediments with A-type ripple-drift cross-laminations that indicate a transport direction of due south. The sloping ridge is interpreted to be an esker formed by meltwater flowing southwest in a subglacial tunnel. Where the meltwater stream issued from the ice it constructed a large ice-contact delta graded to the level of Lake Hitchcock. The fine-grained lake sediments on the east flank were formed by south-flowing currents in Lake Hitchcock after the ice margin retreated from the head of the delta and the subglacial tunnel was destroyed by melting of the ice.

Proceed east on Tyler Hill Road

13.2 Turn left on Route 142. Proceed north on stream terraces cut by the Connecticut River

15.2 Road descends to the level of the Connecticut River

16.5 Turn left with caution just beyond last building on the left at Cersosimo Lumber Company and park adjacent to Route 142. Walk southwest to exposed sediments

STOP 3. CERSOSIMO LUMBER COMPANY. The site is 2.9 km (1.8 mi) S11E of the junction of Routes 142 and U.S. 5 in downtown Brattleboro (Brattleboro, Vt-NH, metric quadrangle). The exposure extends north-south along the west side of the Cersosimo Lumber Company and faces the Connecticut River about 300 m to the east. Two terrace levels can be seen at the top of the exposure. The north terrace is underlain by 2.9 m of pebble gravel and the south terrace has 2.5 m of pebble gravel capped by 0.9 m of very fine sand. Bedrock is exposed 65 m west of the edge of the north terrace and near Route 142 where the cars are parked.

Under the stream-terrace deposits at the south terrace are 5 m of ice-contact deposits that have been disturbed. A distinctive massive bed with vertical fluting is about 1.5 m thick and consists mainly of medium to coarse sand with granules and small pebbles. The bed has lost its original layering and instead displays a crude banding. At first impression, it appears that the layer represents a grain flow, but no channeling is observed and practically all the other beds show similar internal disruption, including the beds under the north-terrace deposits.

At the north under the stream-terrace deposits are 19.7 m of collapsed ice-contact deposits consisting of pebble gravel, sand and silt in disrupted beds. Lenses of interbedded sand and pebble gravel, flame-like structures of pebble gravel extending upward into massive sand, and zones of silt "clasts" formed by dismemberment of original silt layers all appear to be manifestations of massive disruption of bedded ice-contact deposits. Pebble gravel with cobbles is exposed at the north end of the exposure. The disrupted beds are interpreted to have been originally formed in a subglacial stream environment and to have undergone internal deformation caused by massive collapse by melting of adjacent ice and/or some other process.

Proceed north on Route 142

18.5 Stop sign, turn left and sharp right on U.S. 5. Proceed north through the center of Brattleboro. From MILE 20.0 to 21.0 is the Brattleboro "strip" with many fueling and dewatering possibilities on a major stream terrace

21.0 Cross Route 9, proceed north on U.S. 5

21.6 At the traffic light turn right (east) on Ferry Road

21.8 The open area on the right now occupied by United Parcel Service, baked-goods distributor and C & S Grocery is the site of a former ice-contact delta (a morphosequence) that has been removed almost completely by man

22.2 A partially excavated ice-contact delta (another morphosequence) with foreset bedding exposed is at the left. The deposit formerly covered the lumber yard area

22.4 Turn left (west) opposite asphalt plant into Allard pit.

STOP 4. ALLARD PIT: Near the southeast corner of the Town of Dummerston, Vermont, the pit is 2.3 km (1.4 mi) N45E of the junction of Routes 9 and U.S. 5 at the north end of Brattleboro. The maximum elevation in this pit is over 126 m on the Newfane, Vt-NH, 7.5' x 15' metric quadrangle.

A remnant of the highest part of the landform underlies a solitary pine tree in the northern part of the pit. Under the pine tree are thin flat-lying gravels interpreted to be deltaic topsets below which are foresets that dip to the southwest. Good exposures of foresets are northwest and east of the pine tree. The foresets on the east grade southward into a spectacular exposure of collapsed bottomset beds. Bottomset beds located southwest of the pine tree are not collapsed and ripple-drift cross-lamination in fine sand is well displayed. Just northeast of the pine tree is pebble gravel in fluvial crossbeds 60 cm thick that are interpreted to have been formed in a subglacial tunnel. The deltaic deposits are inferred to have been formed in Lake Hitchcock when the ice margin was located at the north end of the pit area. The elevation of the topset-foreset con-

tact was measured to be 129.8 m (425.87 ft) ASL, which is below "the water plane" of Koteff and Larsen (1988).

Return to Ferry Road and proceed west

23.6 Turn right (north) on U.S. 5. Just north of I-91 are lake-bottom deposits of Lake Hitchcock. The road crosses several bedrock ridges and at Hidden Acres Campground enters an area of collapsed stratified drift with many depressions and ridges

26.0 Turn sharp right on dirt road. Proceed east over gravel ridges (eskers) mantled with lacustrine sediments. A former ice margin position was located 200 to 300 m south of and parallel to the road

26.4 Small borrow pit on the right has thick varves (12 to 50 cm) deposited on the north flank of an ice-contact delta

26.5 "Tornado-like" wind associated with a severe thunderstorm of July 14, 1988, demolished a house trailer at the right opposite the former Moore Farm

26.6 Pass under I-91, bear right and drive up into pit

STOP 5. SIMEONE PIT. The pit is located adjacent to the Connecticut River 4.2 km (2.6 mi) N35E of the junction of Routes 9 and U.S. 5 at the north end of Brattleboro (Newfane metric quadrangle, 1:25,000). The landform is the remnant of an ice-contact delta built directly into Lake Hitchcock. Lake-bottom deposits just west of I-91 indicate that the west side of the delta represents part of the original ice-contact slope. The elevation of the topset-foreset contact was measured to be 127.7m (419 ft) which is below that at the Allard pit (STOP 4) and well below "the water plane" of Koteff and Larsen (1988).

Proceed west

27.9 Turn right proceed north on U.S. 5

28.2 The KOA Campground is on a post-Lake Hitchcock meteoric delta built eastward out onto Lake Hitchcock varved sediment. The road drops into the Salmon Brook valley cut in varved silt and clay, rises onto the same delta (fan) surface as at KOA, and gradually descends to lake-bottom deposits at the Ranney Farm, on the right with four blue silos

29.5 Park on right side of U.S. 5 and walk south and east to clay pit owned by Robert Ranney

STOP 6. CANOE BROOK SECTION. The clay pit is located on the north side of Canoe Brook in the Town of Dummerston and is 2.15 km (1.35 mi) N42E of East Dummerston. The semi-circular pit is

about 51 m in diameter and faces south-southwest. The section consists of 21.0 m of thin varves capped by 3.0 m of artificial fill from excavations on I-91. The varves consist of clay-silt couplets 1.0 to 10 cm thick interbedded with fine to very fine sand. Ripple-drift cross-lamination in fine sand dips eastward suggesting a source in the Canoe Brook valley.

On June 29, 1988, Jack Ridge and Fred Larsen counted 640 couplets and discovered brown peat-like organic debris that has been submitted for carbon-14 dating. We hope that a stratigraphic curve showing secular variation of paleomagnetic variation, to be measured by Ridge, plus the carbon-14 date will permit age correlation of Lake Hitchcock varved sediment throughout the Connecticut Valley.

Proceed north on U.S. 5

30.9 Turn right just north of Sunoco station, enter I-91 southbound

32.6 I-91 rises onto stream terrace deposits of the Connecticut River east of the Ranney Farm

34.4 Moore Farm Road passes under I-91

34.5 Simeone pit (STOP 5) on left

35.8 Allard pit (STOP 4) on left

36.0 Top of small ice-contact delta

37.1 Exit right from I-91

37.6 Traffic lights, proceed straight ahead (east) across U.S. 5 and enter Route 9

37.9 Connecticut River with outcrops at river level. The road climbs the flank of a delta on the right and follows the till-stratified contact nearly to the Chesterfield Inn which is on till

40.4 Turn left (north) on Brook Street

41.0 Stop sign, turn left on Main Street which becomes River Road north of the village of West Chesterfield. The road descends a series of small stair-like terraces to the flood plain of the Connecticut River

41.8 Landform west of the Connecticut River is the ice-contact delta of STOP 5

43.3 Park on right at entrance to pit. Just east of River Road are varved silt and clay capped by stream-terrace deposits of the Connecticut River

STOP 7. THOMAS PIT: The pit is located on the south side of the Chesterfield-Westmoreland town line about 350 m east of the Connecticut River on the Newfane, Vt-NH, 7.5' x 15' quadrangle. The pit is in the central part of a landform that is 1.1 km long and 0.45 km wide. The easternmost face has 2 m of topset beds on top of 13 m of southeast- and southwest-dipping foreset beds. The topset-foreset contact was measured to be 131.7 m (432 ft). The projected level of Lake Hitchcock at this site according to Koteff and Larsen (1988) is about 136 m (446 ft).

About 90 m north of the Thomas pit is the Gesmaldi pit with an active pit face about 12.8 m high with 2 to 3 m of topset beds of pebble gravel over foreset beds of sand, pebbly sand, and pebble gravel that dip to the southeast.

The landform is interpreted to be an ice-contact delta that formed directly in Lake Hitchcock between ice on the west and the valley wall on the east. It appears to have been fed, at least in part, by a subglacial stream that also formed an esker, now partially buried by lacustrine sediments about 1 km north-northwest of the Gesmaldi pit.

Extending 1.7 km north of the Gesmaldi pit is an area of post-Lake Hitchcock sand dunes up to 300 m wide. Practically all of the dune area lies below the projected level of Lake Hitchcock. In map pattern the dune sand rests mostly on bottom sediments of Lake Hitchcock and is 5.5 or more meters thick.

Return to cars and proceed north on River Road

43.4 Small pit on right has stream-terrace deposits on lake-bottom sediments on esker gravel

43.7 Coyote Canyon Road on right leads to Gesmaldi pit

44.2 Ridge on right is probable esker covered by Lake Hitchcock bottom deposits

44.5 Small pit at left has 3 m of typical stream-terrace deposits consisting of yellowish-brown fine sand

44.9 Turn right into driveway of white house (Blood residence)

STOP 8. BLOOD PIT: Located in the southwest part of the Town of Westmoreland, New Hampshire, the pit is 3.6 km (2.3 mi) N58E of East Dummerston, Vermont, on the Newfane, Vt-NH, 7.5' x 15' quadrangle. Watch for Mr. Blood's tame partridge.

The pit is in postglacial talus with angular blocks of slatey phyllite probably derived from the Devonian Littleton Formation. Some erratics (possible Cheshire quartzite?) and diamict occur under 3 to 4 m of talus at the base. Bedrock is exposed in the middle and upper parts of the borrow area. The projected level of Lake Hitchcock extends across the upper part of the face. Possibly, the diamict accumulated as subaqueous slide breccia before Lake Hitchcock drained, and the talus accumulated after the lake drained.

From the south end of the pit walk west down gentle slope to base of steeper east-facing slope, 2 to 3 m high. Note lack of stream at base of slope. Climb slope and proceed north and northeast along crest of dunes to small borrow area next to pit-access road. The orientation of sand dunes and wind-abraded bedrock elsewhere in New England suggests that geologically effective winds were from the west and northwest in late-glacial and postglacial time. Canoe Meadows, an extensive terrace cut in fine-grained sediments of Lake Hitchcock, lies immediately to the northwest and probably constitutes the source area for the sand dunes.

45.5 Turn right and proceed north on River Road along the southeast margin of Canoe Meadow, a major stream terrace of the Connecticut River
46.8 Enter Keene metric quadrangle (MAP 5)
48.3 Park in borrow area at right

STOP 9. CHESHIRE COUNTY PIT. The site is located 2.7 km (1.7 mi) N75W of the village of Westmoreland on the Keene, NH-Vt, 7.5' x 15' metric quadrangle. The site offers a mixed bag: (A) from 100 to 150 m northeast along River Road is very compact, probable lower till with iron oxide stain on the joint surfaces, (B) 80 m southwest along the road is a small exposure of varved silt and clay, and (C) in the southwest part of the pit is a large striated erratic of Ascutney syenite, the source for which is located 55 km toward N5E.

Proceed northeast on River Road
49.0 Cheshire County prison on left. After crossing the brook, the road rises to a small terrace then turns 90 degrees to the right and climbs a slope underlain by very compact, oxidized till
49.7 Yield sign, turn left on Route 63
50.1 Park at borrow area on the right

STOP 10. SAND DUNE AREA. The site is on the east side of Route 63, 2.6 km (1.6 mi) N20W of the village of Westmoreland. As at STOP 9, this area is located southeast of a major stream terrace of the Connecticut River, which seems to be a favorite locus for the accumulation of wind-blown sand following the lowering of Lake Hitchcock. The section from the top down consists of: (1) eolian fine sand with buried soil profile, (2) stream-terrace deposits with pebble lag at the base, (3) proximal lacustrine sand with ice-rafted clasts(?), and (4) pebble gravel or till

* To reach Keene, proceed north on Route 63 1.7 miles to Route 12, turn right (southeast) to Keene

* To reach Brattleboro, retrace route on River Road or proceed south on Route 63 to its junction with Route 9, turn right (west) to Brattleboro

# STRATIGRAPHY, STRUCTURAL GEOLOGY AND THERMOCHRONOLOGY OF THE NORTHERN BERKSHIRE MASSIF AND SOUTHERN GREEN MOUNTAINS

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## ROAD LOG FOR TRIP B-1

(This road log is a continuation of field trip A-1, see text of that trip and figures contained for explanatory material)

Assembly point: Parking lot of Grand Union 1/2 mi. east of Wilmington, VT, on south side of Rt. 30.

Assembly time: 8:30 am, Saturday, October 15, 1988

Mileage (cumulative)

0.0 Turn west (left) on Rt. 9 and drive through Wilmington

2.4 Large roadcut of Wilmington Gneiss of Skehan (1972), consisting of strongly retrograded biotite-quartz-plagioclase gneiss having a strongly developed gently north-east dipping Paleozoic foliation. This is typical of one of the gneiss units (Ybg) exposed in the basement-cored strongly-westward overturned F1 antiforms in the Wilmington and Sadagwa Pond areas. An Acadian, biotite, Ar/Ar cooling age has been obtained from this outcrop (GM29 in figure 11). This rock is unconformably overlain on the slopes to the north by albitic cover sequence rocks, quartzite, and dolomite, to be seen at Stops 1 and 2.

3.2 Turn left onto bridge over river and turn right (west) on New England Power Rd.

3.5 Turn left behind large barn, consolidate into 4-wheel drive vehicles for drive 0.5 miles up hill to small cabin and park (log resumes on return).

STOP 1. Medburyville occurrence of marble described by Skehan (1961) and basal cover sequence of the western fland of the Wilmington antiform.

Walk from cabin east to powerline to exposures of dolomitic marble and albitic granofels of the Hoosac Formation. Just east of the powerline are aseries of low outcrops of typical albitic Hoosac conglomerate, having a ghost-like gneissic inclusions in an albitic matrix.

4.2 From barns retrace route to Rt. 9 and turn east (right).

7.8 Intersection of Rt. 100, Wilmington, turn left and continue north through West Dover. The route follows the core rocks of the antiform and a major thrust fault in the gneiss. To the west are Haystack Mountain and Mount Snow which consist of cover rocks of the western flank of the Wilmington antiform.

19.7 Entrance to Mt. Snow Ski Area, turn left from Rt. 100.

20.0 Stop sign turn right then left and part at maintenance shed.

20.5 STOP 2. Mount Snow traverse - Hoosac cover rocks

The upper Proterozoic and Lower Cambrian cover sequence rocks on the westside of the Middle Proterozoic Wilmington Gneiss belt of Skehan (1972) consist of a fairly regular succession that has been mapped from Mount Snow south to Hoosac Mountain (see fig. 6). Basal albitic conglomerate (like at Stop 1), feldspathic quartz pebble conglomerate like the Dalton Formation, or albitic granofels unconformably overlie the basement rocks. In the Mount Snow area, a succession of conglomerate and albitic granofels, having discontinuous lens of biotite-actinolite greenstone and beige or salmon-pink weathering dolostone form the basal section. Above this a discontinuous large-garnet-muscovite-biotite-plagioclase quartz schist or chlorite-garnet-chloritoid-muscovite quartz schist is found (CZhgt). Immediately above the garnet schist or locally directly above the albitic granofels and dolomite marble, is a thick succession of dark-gray to


Figure 15. Generalized geologic map of the souhern part of Green Mountains massif and Sadawga-Rayponda domes area. Solid dots show field trip stops. Solid triangles show location of 40Ar/39Ar samples discussed, numbers refer to figure 18. For explanation of rock units see figure 1, Trip A-1.

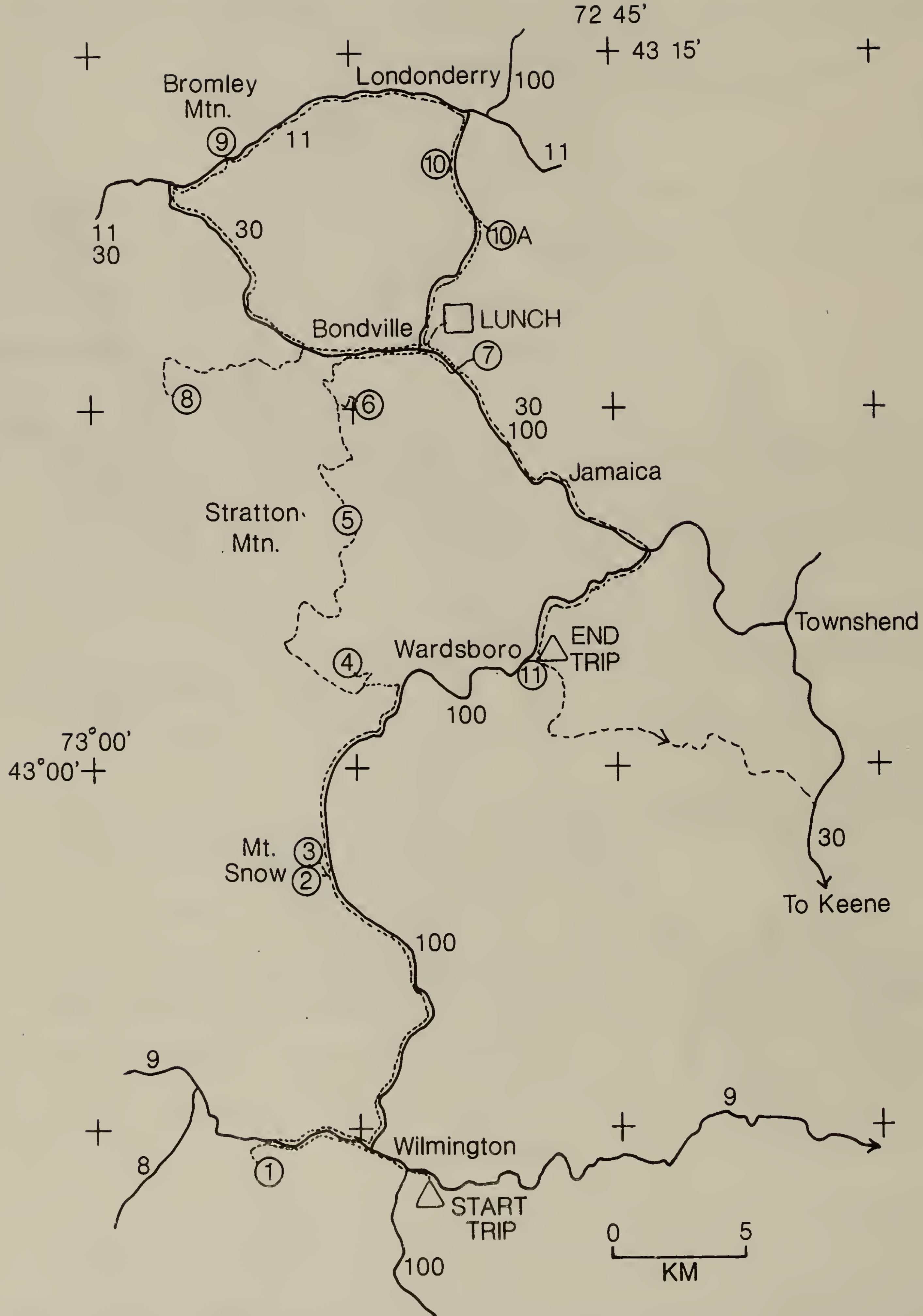


Figure 16. Route of field trip stops and stop locations for trip B-1.

black, very fine-grained, phyllitic biotite±garnet-quartz-albite schist that locally pass into a dark chloritoid-bearing phyllite. This unit (CZhgb) forms the bulk of the Hoosac Formation on Mount Snow. Above this a second aluminous garnet schist is found (CZhgt).

We will begin our traverse at the 840 m level on the northside of the mountain where the dolomite marble unit is well exposed, and walk down the ski slopes across the section down to the lower albitic beds.

Structures in the Mount Snow area consist of northwest to southeast F1 folds, having north-dipping axial surfaces. These folds are responsible for the distribution of the major map unit. F2 folds strike north to northwest, dip southeast and are spatially associated with a thrust fault exposed to the west at the Searsburg thrust, and to the east at the Wilmington thrust. Post-thrust crenulation cleavage trending NE is subvertical, and a fourth cleavage trending NNW produces open cross folds and local crenulation cleavage.

Return to base of slope and walk 600 feet north across access road to exposures of mylonitic Kspar megacrystic gneiss

STOP 3. These excellent exposures of biotite-rich augen gneiss are representative of the coarse megacrystic granite in the core of the Wilmington antiform above the Wilmington fault system.

Early-generation fold (F1) structures in the Mount Snow belt are truncated by this fault. This gneiss is probably equated with the 958 Ma old Stamford Granite Gneiss.

Return to cars. Exit Mt. Snow Ski Area via same route.

21.2 Turn left (north) on Rt. 100

21.8 On left side of road excellent crops of arkosic conglomerate and gritty rocks that unconformably overlie the coarse Kspar megacrystic granite and other gneisses above the Wilmington fault. The unit forms the base of a somewhat different succession of Hoosac Formation and Turkey Mountain member (of Doll and others, 1961) from that in the Mount Snow belt.

23.7 Exposures of garnet-chlorite-muscovite-chloritoid quartz schist (CZhg) west of the Wilmington fault, enter Jamaica quadrangle.

26.6 At base of long hill turn left onto road to Stratton.

27.6 Turn right into Mt. Farms West, turn right immediately and follow road curving to left up hill to 4-way intersection, turn left and follow curving road to the right up the hill slope near crest of hill at roadcuts.

29.4 STOP 4. Kspar megacrystic granite and mylonitic gneiss (Stratton Mtn. quadrangle)

This coarse-grained granite is traceable to the southwest into a wide belt of biotite rapakivi granite like the Stamford Granite and traceable eastward into biotite augen gneiss in the bed of Wardsboro Brook referred to as the Bull Hill Gneiss by Doll and others (1961), and dated by zircon U/Pb technique by Paul Karabinos (Karabinos and Aleinikoff, 1988) as 950 Ma. This granite is unconformably overlain by rusty albitic schists of Hoosac Formation that contain distinct beds of feldspathic quartzose conglomerate identical to the basal beds of the Dalton Formation (CZdsc) seen at Stop 1 on Trip A-1. Clearly this belt of Bull Hill-like gneiss is a part of the Green Mountain massif basement rocks.

Follow loop of road down to base of hill and exit right.

30.7 Onto paved road to Stratton

33.1 Turn right on West Jamaica Rd.

35.4 Turn left on Mountain Road headed toward Stratton Ski area

36.2 Begin outcrops of well-layered biotite-quartz-plagioclase gneiss and rusty schistose gneiss. Continue north past Forrester Road and stop near crest of hill at large freshly blasted outcrops.

38.1 STOP 5. Migmatitic gneiss and pegmatite (Ymg) on Stratton Mountain

These excellent roadcuts of migmatitic 2-feldspar-biotite granite gneiss (Ybm) are typical of one type of syntectonic granitoid rocks present in the Green Mountains. In the Jamaica-Stratton Mountain area several belts of these rocks have been mapped. One belt forms a mantle or wreath around a coarse Kspar megacrystic granite exposed at College Hill (Stop 6). These gneisses are high SiO2, K2O rich rocks having low CaO (table 4). In figure 4 they are grouped with the Kspar-rich granitic rocks (Ygg) that lack the abundant relict structures and short-like gneissic lens in these rocks.

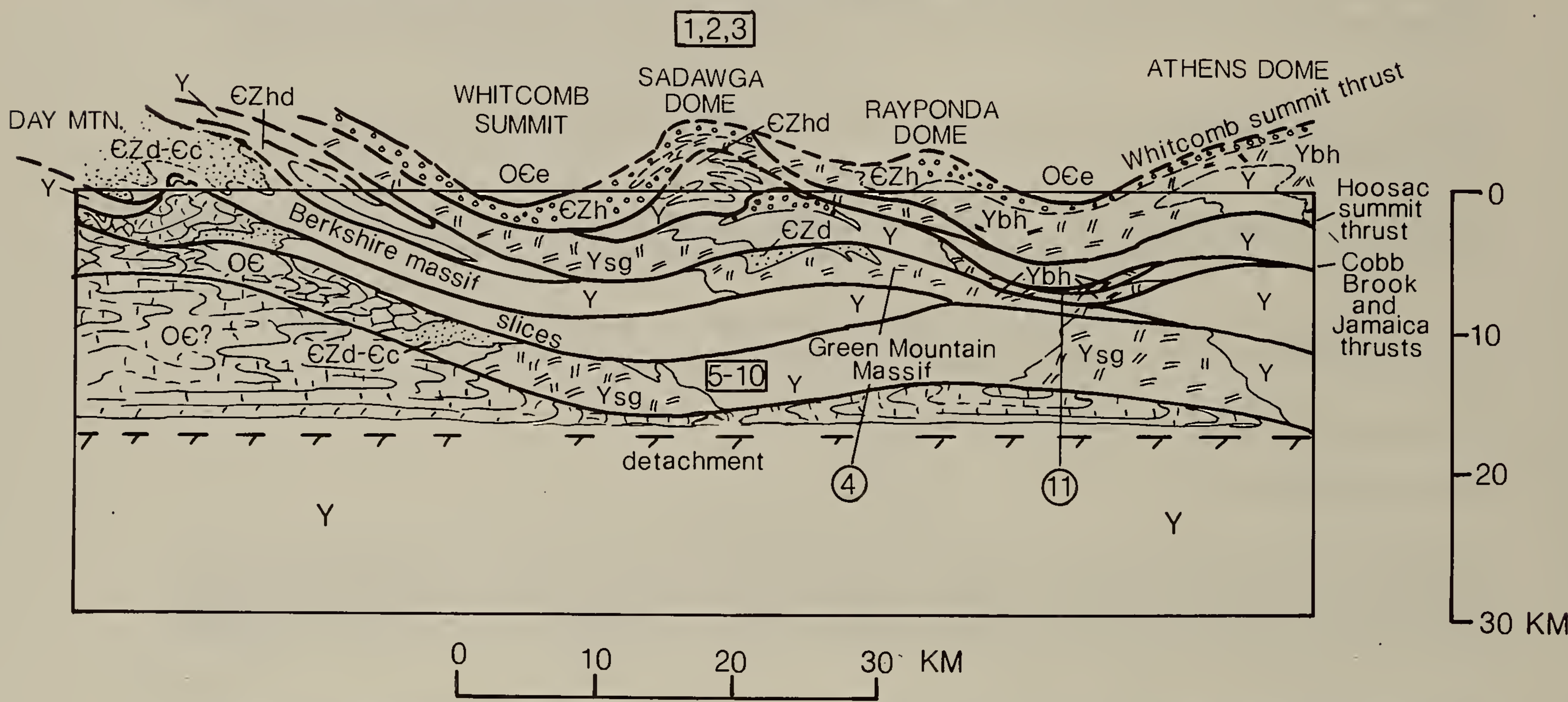


Figure 17. Schematic cross section from start of trip A-1 at Day Mountain, Massachusetts northeast to the Athens dome showing correlation of major faults and lithotectonic units. Numbers refer to projected location of selected field trip stops, Trip B-1.

Paleozoic foliation trends N 40° E and dips SE parallel to axial surfaces of late folds. Hornblende from Stratton Mountain (1178A) produced a disturbed Ar/Ar release spectra characteristic of Proterozoic Y hornblendes in zones of biotite-grade Paleozoic remetamorphism (fig. 18).

Continue N on road to T

40.6 Intersection and turn right on Work Rd.

41.4 Turn left at base of hill onto Pikes Falls Rd., continue north over Ball Mtn. Brook to....

42.7 The third right turn, and turn onto Benson Fuller Road. Bear right at Y onto Pearl Buck Drive

43.4 Stop near crest of hill

STOP 6. Biotite megacrystic granite of College Hill (Londonderry quadrangle).

Abundant exposures of Kspar megacrystic biotite granite gneiss having Proterozoic Y deformational structure older than coarse biotite granite pegmatite. Granite like this forms a prominent unit, bordered by migmatitic gneiss. This rock resembles closely in chemistry and overall appearance the Tyringham Gneiss of the Berkshire massif (table 4 and fig. 4). This very distinctive megacrystic gneiss can

be traced (fig. 5) from College Hill westward across the north slopes of Stratton Mountain into the Sunderland quadrangle thus ruling out a major fault in this area despite strong mylonitic fabrics.

44.3 Proceed to end of turn-around and return to Pikes Falls Road.

45.1 Turn right (north) and go to intersection with Stratton Village Road, turn right and follow to intersection with

46.6 Rt. 30.

49.3 Road, turn left

50.0 STOP 7. Biotite tonalite gneiss at Cole Pond (Londonderry quadrangle).

Distinctive, biotite-quartz-plagioclase metatonalite, biotite-poor-plagioclase-quartz (metadacite), biotite-spotted granodiorite, and white, fine-grained plagioclase-rich aplite rocks form a distinctive group of rocks in the Peru, Londonderry, Stratton Mountain, and Jamaica quadrangles. Coarse-grained rocks are massive, nonlayered to weakly foliated rocks suggestive of metaigneous rocks. Exposures here are typical of the mafic, biotite-rich members of this group of rocks. Together with locally devloped plagioclase hornblende gneiss and amphibolite these rocks may constitute a suite of metaintrusive and volcanic calcic to calc-alkaline rocks not recognized in such abundance elsewhere in the Berkshire massif or southern Green Mountains. Collectively these rocks appear to underlie or to intrude the well-layered biotite-quartz-plagioclase paragneiss, quartzite, and calc-silicate rocks that form the bulk of the core rocks to the south. Chemically these rocks resemble rocks of the Losee Metamorphic Suite of the Reading Prong, the biotite-quartz-plagioclase leucogneiss of the Hudson Highlands of New York and certain diorite to trondhjemite rocks associated with these rocks (see table 6 and fig. 7). Their presence in the Green Mountains has just recently been recognized. They may correlate with certain tonalitic gneisses in the eastern Adirondacks recently recognized by McLelland to have U-Pb zircon ages possibly as old as 1.3 Ga.

50.9 Return to Rt. 30 turn right (west)

51.9 Turn right on Goodalville Rd., in 0.6 mi bear right just before bridge over Winhall River and park at slight bend in road.

Lunch stop. Mylonitic white aplitic gneiss, and trondhjemite exposures in brook, located along the trace of the Londonderry-Stratton Mountain fault system.

53.5 Return to Rt. 30, turn right and proceed.

56.6 Turn left on Kendall Farm Rd., leaving Winhall Memorial Library to port (left), follow dirt road to point just past large meadow.

58.6 Turn left onto unmarked road and drive 0.3 mi to gated entrance to U. S. Forest Service road. (This gate is normally locked and closed to vehicular traffic). Follow dirt road uphill over low outcrops of medium to coarse-grained biotite trondhjemite or tonalitic gneiss.

60.7 Several large boulders of typical biotite trondhjemite gneiss

62.2 STOP 8. Paragneiss and amphibolite structurally above the metatrondhjemite-metatonalite unit.

Rusty weathering, biotite-garnet schists, biotite-magnetite gneiss, hornblende-garnet amphibolite. Pavement exposures of well-layered gneiss constituting the southwestern border of the trondhjemite-tonalite-aplite gneiss belt in the Peru quadrangle. Coarse garnet-biotite-plagioclase quartz gneisses and rusty garnet-quartz-sillimanite(?) gneisses. Walk to exposure of hornblende gneiss, diopside-hornblende sulfidic rocks 100 feet to the south. A succession very similar to this marks the border of the white gneiss belt throughout this area. Where intruded by white pegmatite and retrograded in the Paleozoic, lustrous, pale-green chlorite-muscovite quartz$\pm$chloritoid phyllites are developed from these rocks. Excellent exposures on the Pinnacle in the Londonderry and Jamaica quadrangles, and the Peak of Stratton

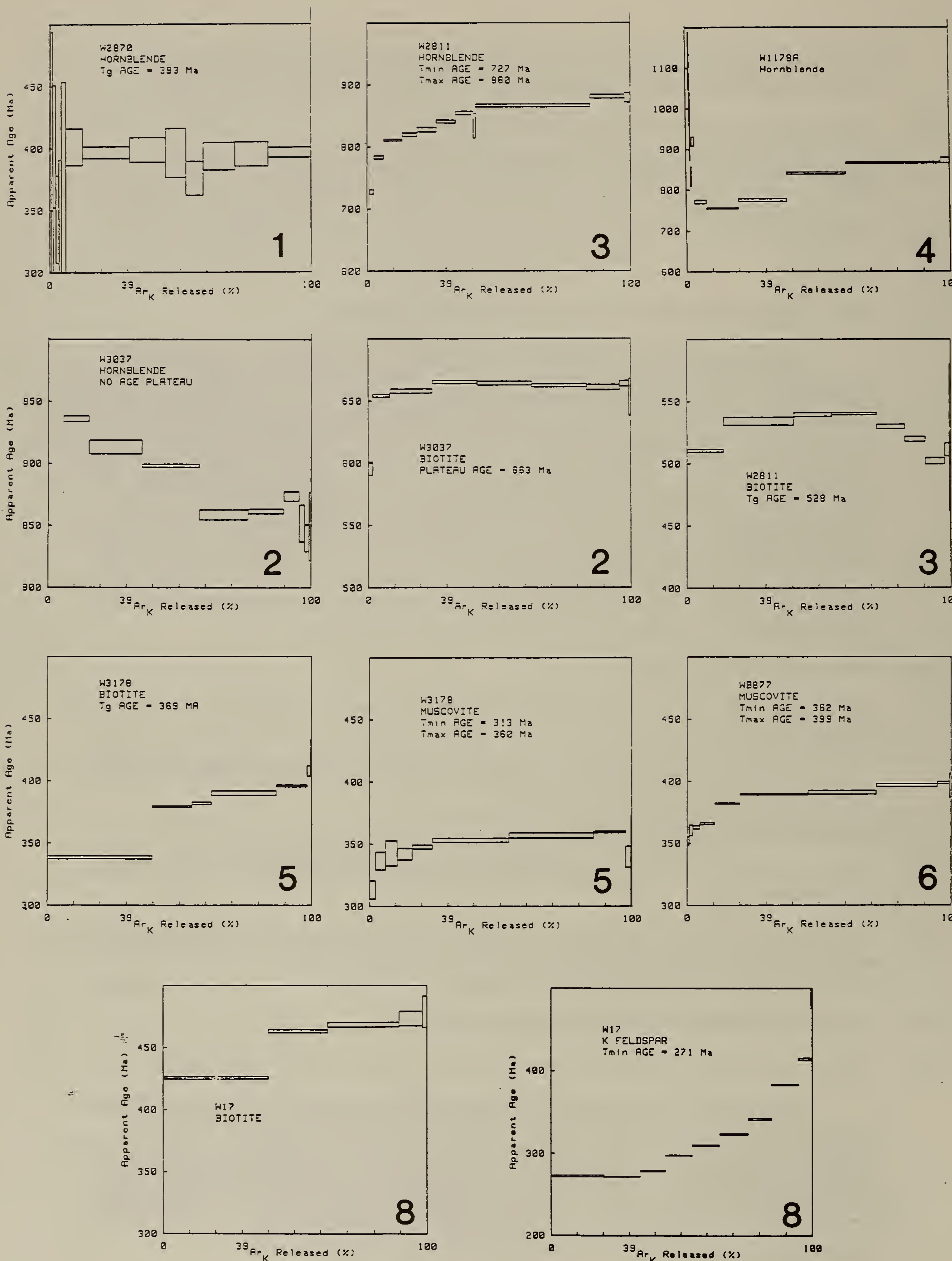


Figure 18. Selected $^{40}Ar/^{39}Ar$ age spectra from the southern and central Green Mountains, Vermont. Numbers refer to locations indicated on figure 15 (location 7 data not shown). See text for discussion.

## $^{40}Ar/^{39}Ar$ THERMOCHRONOLOGY

The area of trip B-1 is entirely north of the Green Mountains traverse of Mukasa discussed in trip A-1. As noted before, the central Green Mountain massif in the vicinity of this trip is at least in part in the biotite zone of Paleozoic (presumably Taconian) regional metamorphism. At a point near Jamaica, VT, coarse-grained garnet-muscovite-chlorite-quartz schists are infolded with basement gneisses, for example, in the easternmost cover rock synform shown in figure 15. Karabinos (1984) has identified the staurolite isograd 2.5 km southeast of Jamaica, just outside the region of figure 15. Crystalloblastic textures and biotite and hornblende 40Ar/39Ar closure ages indicate the eastern, higher-grade rocks enjoyed Acadian metamorphism at about 380 Ma (Karabinos, 1984; Sutter and others, 1985). The area west of the staurolite isograd undoubtedly contains effects of both Taconian and Acadian metamorphism.

In an attempt to delineate these zones, a collection of hornblende, muscovite, biotite, and microcline-bearing assemblages was made for 40Ar/39Ar thermochronology from basement and cover rocks west of the inferred Paleozoic garnet isograd. The preliminary results (age spectra) are shown in figure 18.

Hornblende-bearing samples from Middle Proterozoic gneiss listed by increasing Paleozoic grade are: W1178A (no. 4 in figs. 15 and 18; near stop 5), W3037 (no. 2; stop 10A), W2811 (no. 3; near stop 6), and W2870 (no. 1). The first three of these hornblendes yield discordant age spectra characteristic of variably disturbed Middle Proterozoic amphiboles from other areas in the Green Mountain massif (Sutter and others, 1985). None of these hornblende age spectra forms an age plateau but using these and other published spectra we can make a case for initial cooling to argon closure in hornblende about 900 million years ago, followed by a thermal disturbance, probably Paleozoic metamorphism. Biotite from two fo these samples, W3037 and W2811 (no. 2 and 3 respectively in figs. 15 and 18) yield variable discordant age spectra, with W2811 biotite yielding a "total gas" age of 528 Ma and W3037 biotite defining an "age plateau" at 663 Ma. Despite the age plateau of W3037, we are hesitant to assign any geologic meaning to these apparent ages because they are similar in character to ages and age spectra for incompletely retrograded Proterozoic gneisses in the eastern Reading Prong (Dallmeyer and Sutter, 1976). These data do suggest that neither Taconian or Acadian thermal effects were sufficient in this area to totally reset even the K-Ar system in biotite (about 300°C). However, the amphibole from sample 2870, collected from a typical by minor thrust fault near the eastern margin of the massif at Paleozoic garnet grade, yields an only slightly discordant age spectrum in which nearly 81% of the total argon defines an age of 396 Ma. Unlike the other hornblende-bearing samples, textures in W2870 indicate recrystallization in a mylonitic fabric associated with Paleozoic ductile fold-thrust structures. This apparent age could be interpreted either as the result of very slow Taconian cooling (2-4°C/million years; Sutter and others, 1985) or as the time of formation during Acadian recrystallization. The interpretation rests heavily on the temperature maximum during the mylonitic fabric development (in this case garnet grade) and the subsequent cooling rate.

The age spectra of muscovite from sample WB877 (no. 6 in figs. 15 and 18) and of biotite and microcline from W17 (no. 8) are all compatible with cooling from the Taconian metamorphic maximum. Sample WB877, collected near the western margin of the massif from the Dalton Formation, yields a plateau age of about 390 Ma. This is the first 40Ar/39Ar age from muscovite in the cover rocks of the massif and is very significant since it suggests that temperatures associated with Taconian metamorphism of the T-2 zone (fig. 10, trip A-1) reached closure temperature for muscovite about 390-400 million years ago. Biotite from W17, mylonitic Stamford Granite Gneiss, yields a total-gas age of about 450 Ma consistent with previous results from the area as discussed on trip A-1. Microcline from this sample, W17, shows a steady increase in age with temperature from about 270 Ma to well over 400 Ma. We interpret this age gradient as a result of slow cooling in the temperature range of 150-250°C or even lower. Because there is no evidence in the age spectrum for "extraneous" 40Ar, the results can be used to say it is unlikely that this part of the Green Mountain massif ever saw temperatures greater than 200-250°C after about 415 million years ago. Hence this area must have been structurally high during Acadian metamorphism, in marked contrast to areas just to the east.

Finally, a muscovite-biotite mineral pair (W3178) was collected from a 3-meter thick, vertical strike-slip or contractional ultramylonite zone cutting the granite at College Hill (no. 5 on figs. 15 and 18). The biotite yields a very discordant age spectrum and a total-gas age of about 370 Ma. The muscovite is much better behaved and yields a near-plateau of about 360 Ma. The location of this sample west of the garnet isograd, and in an area where neither biotite nor hornblende are reset to Paleozoic ages, suggests the muscovite age from this mylonite is a good estimation of the time of faulting. Several N50°E trending vertical, apparently strike-slip faults cut the basement gneisses between College Hill and Jamaica, VT. These faults offset the regular Paleozoic overprint fabrics in the gneisses and are demonstrably late features. Our data suggest these are Acadian faults and therefore indicate Acadian shortening in an E-W or NW-SE direction.

Mountain and north of Bromley in the Peru quadrangle closely resemble cover sequence rocks of the Pinney Hollow or the aluminous schist facies of the Hoosac Formation. These pseudo-cover sequence rocks are abundant within the basement rocks, and special care is needed to distinguish them from *bona fide* cover rocks.

Return to Rt. 30 via the same route used to enter

69.0 Turn left of Rt. 30

76.2 Turn right on Rt. 11

76.7 Turn left into Bromley Ski area and park at lodge

STOP 9. Metatrondhjemite gneiss and paragneiss: Bromley Mountain traverse.

The traverse begins at the summit of Bromley Mountain in medium-grained trondhjemite gneiss containing inclusions of hornblende diorite gneiss. Relationships suggest that the trondhjemite intrudes a more mafic host. Exposures to the south contain calc-silicate rocks and quartzite similar to the rocks seen at Stop 8. Contact relationships between the trondhjemite and the paragneiss have not been determined.

Excellent views to the south show the steep west facing scarp composed of west dipping Cheshire Quartzite and Dalton Formation. Detailed mapping from Bennington north to Manchester, by Burton, shows that both foliation and bedding dip west and are strongly folded by F2 generation north-south folds. Locally, early, more east-west trending folds are present in the cover rocks resulting in areas of complexly refolded folds. The west dipping foliation and bedding gives a deceptively simple impression of the geology. At present significant thrust faults between Cheshire-Dalton and Vermont Valley carbonate rocks are a distinct possibility in several areas but much more work is required.

From the top we will walk down the northern ski slope starting in white weathering east-west trending metatrondhjemite gneiss. Coarse gneissosity folded into subvertical isoclinal Middle Proterozoic folds is abundant. Locally spodumene pegmatite cross cuts the folds. At a marked bench in the ski slope about 600 feet from the crest a strong zone of Paleozoic retrogression and transposition can be seen. In a distance of about 10 meters, coarse metatrondhjemite and diorite layers are transposed into a northeast striking, northwest dipping Paleozoic fabric and the rock transformed into a fine-grained well-foliated flaser gneiss.

This section illustrates beautifully the severe nature of Paleozoic retrogression in the basement rocks. Within zones like this chlorite-sericite-actinolite-epidote are common minerals. Paleozoic folds have been mapped from this point down the mountain. It is no wonder, having seen these relations, that identification of original rock types let alone protoliths is difficult in the Green Mountains. Follow the ski slope down to the end of the chair lift.

Return to cars, turn left on Rt. 11, and continue to intersection of Rt. 100 at Londonderry.

83.5 Turn right on Rt. 100. Outcrops in West River of biotite trondhjemite gneiss, dioritic gneiss, and minor amphibolite trend east-west and are subvertical. The slopes to the east and above the West River are strongly foliated and retrograded gneisses in an east-dipping shear zone and possible fault.

84.9 STOP 10. Fault zone and transposition structure

Thrust fabric in retrograde gneiss West River crossing of Rt. 100 Londonderry quadrangle. Excellent chloritic shear zone fabric in this outcrop dips east at about ten degrees. A belt of similarly sheared rocks can be traced northward to Londonderry and south through the lunch stop to the Winhall River south of Route 30 at Bondville. This zone may constitute a regionally important fault zone that runs diagonally from northeast to southwest across the massif from Londonderry to the exposures at the Dome visited on Trip A-1; however, the metatrondhjemite-metatonalite belt is essentially continuous across this zone. Continue south to South Londonderry bridge over West River.

86.4 Turn left for optional Stop 10 or turn right headed south.

86.9 STOP 10 (Optional).

Dirve 0.5 mi to outcrops along road cuts in amphibolite, and coarse-grained Cole Pond-like tonalite. Ar/Ar hornblende and biotite samples were taken from excellent exposures in the creek. Coarse hornblende and well-twinned oligoclase in the amphibolites define an excellent gneissosity that is overgrown by coarse irregular pods of quartz-plagioclase trondhjemite with only sparingly small amounts of K-feldspar. Petrographically and field observation suggest that the fabric and most folds in the rocks area Proterozoic.

87.4 Return to Rt. 100, turn left and continue on 100 to intersection with Rt. 30 at Rawsonville.

91.3 Turn left on 100. Along this route outcrops on both sides of road for next 5 miles and large cliffs above road at Rawsonville consist of white weathering, coarse to fine-grained, biotite-poor (5-10%), tonalitic gneiss and white, fine-grained biotite-quartz-plagioclase leucogneiss.

95.8 Low crops on left approaching Jamaica of white fine-grained quartz-plagioclase leucogneiss sampled for zircon by Paul Karabinos and John Aleinikoff. Preliminary results suggest that this rock may be 1.3 Ga or older (see Karabinos and Laird, this volume). Continue east through Jamaica over West River and turn right on Rt. 100 S. Rt. 30 continues straight.

99.5 Right on Rt. 100. This route, parallel to the Wardsboro Brook, follows a complex belt of highly faulted K-feldspar megacrystic granite (Bull Hill Gneiss of Doll and others, 1961), intercalated Hoosac Formation, and belts of coarse-grained, large garnet-chlorite-chloritoid-muscovite quartz schist. Our mapping indicates that the northernmost belt of granite is traceable into the granite seen at Stop 4 and is physically within the Green Mountain basement. The southeastern belt of granite and other gneisses passes through Wardsboro and is bounded on its southeast side by a complex fault zone that connects with the Wilmington fault system to the south.

102.3 Low exposures in bank of Wardsboro Brook of K-feldspar megacrystic granite dated by Paul Karabinos (Karabinos and Aleinikoff, 1988) approximately 950 Ma by U-Pb zircon technique. This granite and that of the southern Wardsboro belt are Bull Hill Gneiss according to Doll and others (1961). The recent U-Pb data suggest that all of these megacrystic granites including the Stamford Granite of Hoosac Mountain, at Stamford, VT, near Stratton, in the core of the Wilmington antiform, in the Wardsboro belt and in the Chester and Athens domes are all post-Grenville plutonic rocks about 960 Ma old. The close association of these rocks with cover rocks, in the Chester dome and along the east side of the Green Mountains massif is attributable to thrust faulting and/or tight infolding of cover rocks.

103.9 Town of Wardsboro, stop sign, turn left over Wardsboro Brook.
STOP 11. Mylonitic gneiss along Wilmington fault zone.

Mylonitic K-feldspar megacrystic granite of the Wardsboro belt. Excellent mylonitic granite and augen gneiss is exposed under the bridge. THe orignal texture of this rock may have been similar to the coarse-grained ovoidal or rapakivi granite seen at Stamford, VT (Stop 6, on A-1) or the coarse-grained granite seen at Stop 4 today. The fabric in this 960 Ma old granite is entirely Paleozoic and is associated with the Wilmington thrust fault. At this point well-layered rusty, biotite-quartz-plagioclase paragneiss, quartzite, and amphibolite occur under the thrust fault. The eastern albitic Hoosac, containing numerous, and very abundant, greenstones unconformably overlies this gneiss belt. Mapping in the southeast corner of the Jamaica quadrangle suggests that numerous thrust faults are likely within the eastern cover sequence, as a regular coherent stratigraphic succession is not present. Although this work is not completed, the northern extension of the Whitcomb summit thrust passes between this locality and the town of South Wardsboro, 0.5 mi to the south.

End of trip. Instructions to Keene - Continue up dirt road to South Wardsboro, turn left at T intersection, continue 1 mile, take Y at branch in road and follow this dirt road 8 miles down to the town of Newfane where you will rejoin Rt. 30. Take Rt. 30 to Brattleboro, Rt. 5 N and cross the Connecticut River on Rt. 9 (right turn). Follow Rt. 9 to Keene.

# STRATIGRAPHY AND STRUCTURE OF THE MONADNOCK QUADRANGLE, NEW HAMPSHIRE (Trips B-2 and C-1)

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## Introduction

The layered rocks of the Monadnock quadrangle, New Hampshire, have been mapped (Thompson, 1985) based on a stratigraphic sequence which correlates well with the sequence described by Lyons and Hatch in central New Hampshire and earlier by Moench in western Maine (Hatch et al., 1983). Three distinctive Silurian units separate Silurian Rangeley Formation from the Devonian Littleton Formation. In order from oldest to youngest these are: the Perry Mountain Formation, thinly bedded schist and white quartzite; the Francestown Formation, mainly sulfidic calc-silicate granulite with a subordinate sulfidic schist; and the Warner Formation, consisting of a lower clean calc-silicate granulite and an upper feldspathic granulite. The Rangeley Formation includes sulfidic to gray-weathering gritty schist with calc-silicate pods, granulite beds, quartz-pebble conglomerate lenses near the top, and granulite-matrix conglomerate horizons near the lowest exposed part. The Littleton Formation consists of gray-weathering schists and quartzites, with the proportion of quartzite increasing upwards. Mt. Monadnock itself is held up by quartzite beds in the Littleton Formation which are folded back on themselves to form a thick resistant sequence. Davis (1896) cited Monadnock as the type locality for resistant mountains rising above the general erosion level, thereby coining the term "monadnock" for such isolated peaks worldwide.

The Silurian is represented by a much thinner sequence of rocks along the west edge of the quadrangle, where the Clough Quartzite and local Fitch Formation, probably correlative with the Silurian units described above, overlie Ordovician rocks of the Keene dome. The quadrangle thus straddles a Silurian "tectonic hinge" (Figure 1), which may have behaved as a zone of weakness during Acadian deformation.

Intrusive rocks include the pre- or syn-tectonic (Devonian) Kinsman Granite, syn-tectonic (Devonian) Spaulding Tonalite and related rocks, and post-tectonic (?Mississippian) Fitzwilliam Granite. Granite dikes and microdiorite dikes probably coeval with the Fitzwilliam cut all generations of folds on Mt. Monadnock.

Five phases of Acadian deformation have affected the rocks of the quadrangle (Figure 2). Fold nappes and then thrust faults transported rocks of the Merrimack trough (Rangeley through Littleton Formations) westward across the tectonic hinge onto the thinner Bronson Hill sequence. The rocks were then folded back toward the east in two complicated phases which included mylonization along the short limb of a major backfold, the Beech Hill anticline. The final phase produced folds related to the rise of

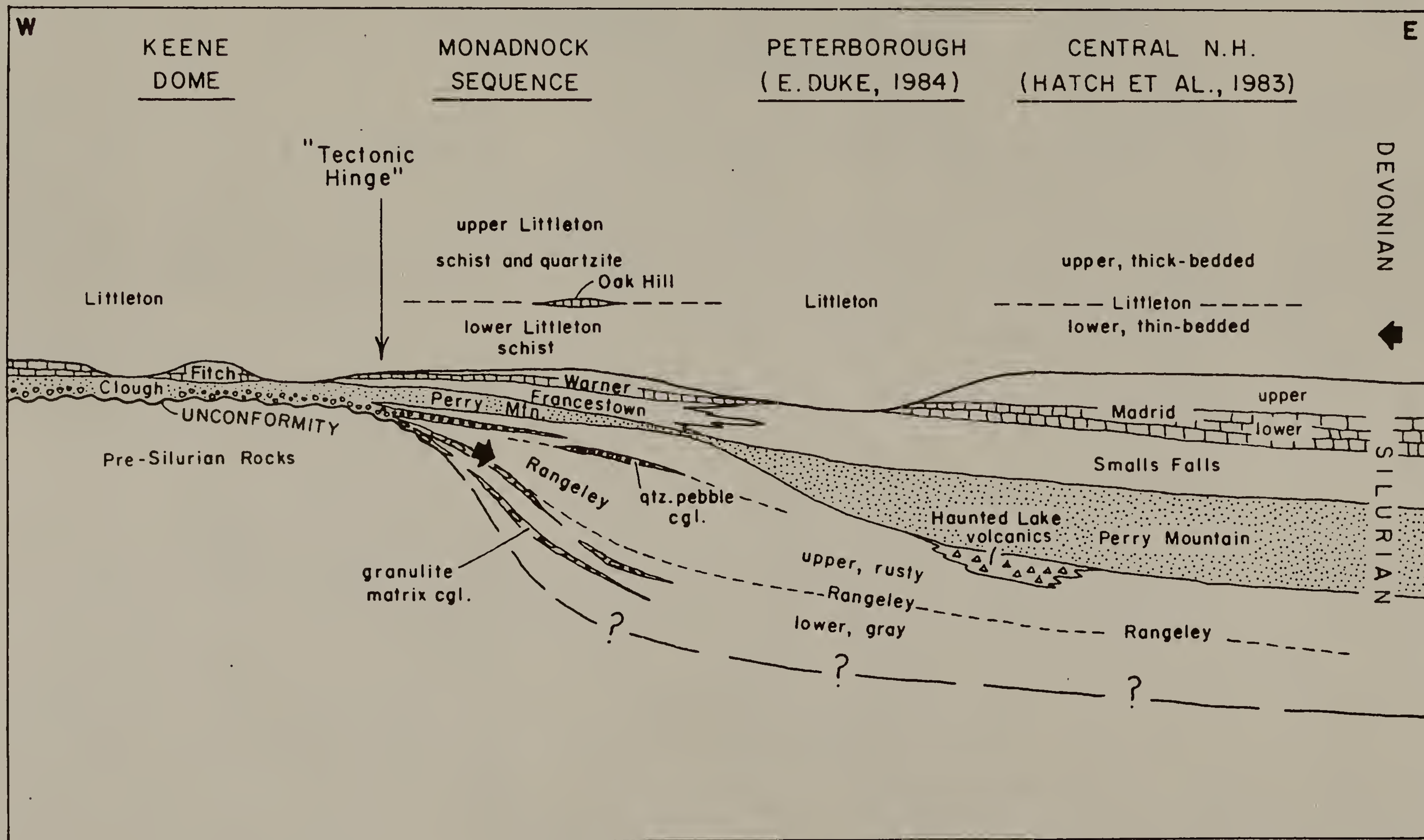



Fig. **1**. Stratigraphic diagram showing proposed correlations across the tectonic hinge, after Hatch et al. (1983), to include the Monadnock and Peterborough sections. Black arrows show change in sediment source direction from Silurian to Devonian.

the Keene dome at the west edge of the quadrangle. There is evidence for several periods of movement along the four-kilometer-wide Thorndike Pond fault zone, including nappe-stage ductile thrusting, east-directed shear during the backfolding, late Paleozoic shear in the Fitzwilliam Granite, and Mesozoic normal faulting and silicification. A single occurrence of a presumably Mesozoic diabase dike was found in a float block of Kinsman Granite.

The dominant foliation lies parallel to axial planes of nappe-stage folds, and the peak metamorphism probably occurred during the early backfolding. Assemblages in pelitic schists range from Zone II in the west to Zone VI near the Kinsman Granite in the northeast, but Zones III (silimanite-biotite-garnet-muscovite) and IV (sillimanite-biotite-garnet-muscovite-K-feldspar) predominate in most of the quadrangle. Garnet and biotite compositions in Zone III yield peak temperature estimates of 635-670° and evidence of re-equilibration during cooling and unloading. A pressure of 6.3 kbar is indicated by garnet composition in equilibrium with cordierite and sillimanite. A zoned calc-silicate pod from Zone III contains bustamite in the core, as well as a variety of calc-silicate minerals with interesting disequilibrium textural relationships.

## STRATIGRAPHY

Fowler-Billings (1949) described most of the layered rocks in the Monadnock quadrangle as members of the Littleton Formation. Her lower and upper schist members are now mapped as Rangeley Formation, and her middle member is retained as Littleton Formation. Her rusty quartzite member is mapped as Francestown and Warner Formations, following Nelson (1975). Topping directions are provided by graded beds in the Rangeley, Perry Mountain and Littleton Formations.

### Rangeley Formation

Rocks assigned to the Rangeley Formation in the Monadnock quadrangle include rusty-weathering sulfidic schists, gray-weathering schists, quartz-feldspar-biotite granulites, a variety of conglomerate horizons and lenses, and a variety of calc-silicate granulite horizons and lenses. The mapping of rusty- and gray-weathering rocks within the Rangeley Formation has not shown any clear internal stratigraphy; the situation seems to be more complicated than simply an upper rusty part and lower gray part, as suggested by Hatch *et al*. (1983) for central New Hampshire. Because of the heterogeneity of the unit, consistent internal stratigraphy should probably not be expected over wide areas.

Three informal parts of the Rangeley have tentatively been identified. The lower part is predominantly gray-weathering schist with calc-silicate pods and local granulite and granulite-matrix conglomerate. The middle part consists of a thick sequence of predominantly sulfidic schist with calc-silicate pods. The upper part consists of interbedded gray and sulfidic

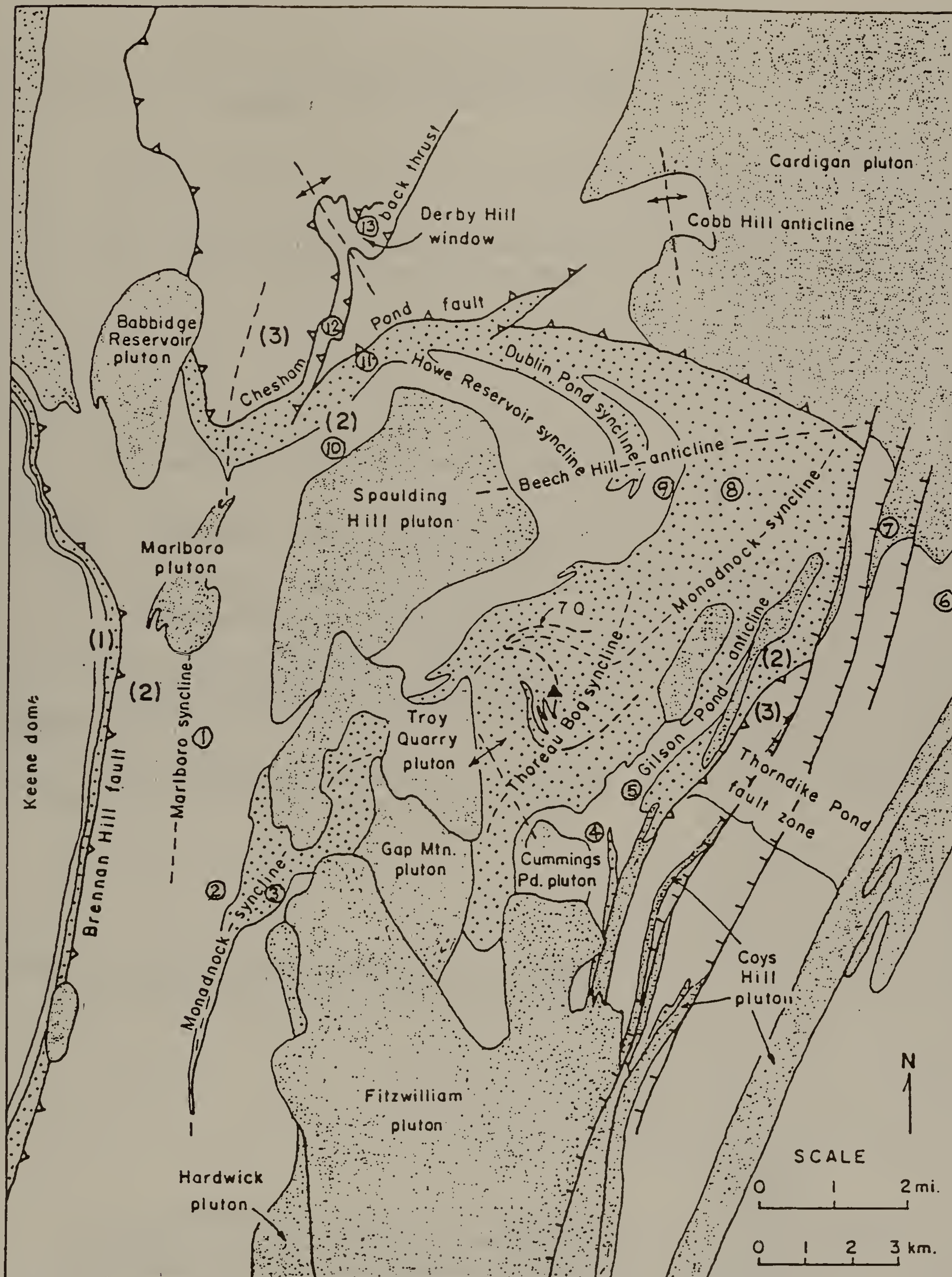
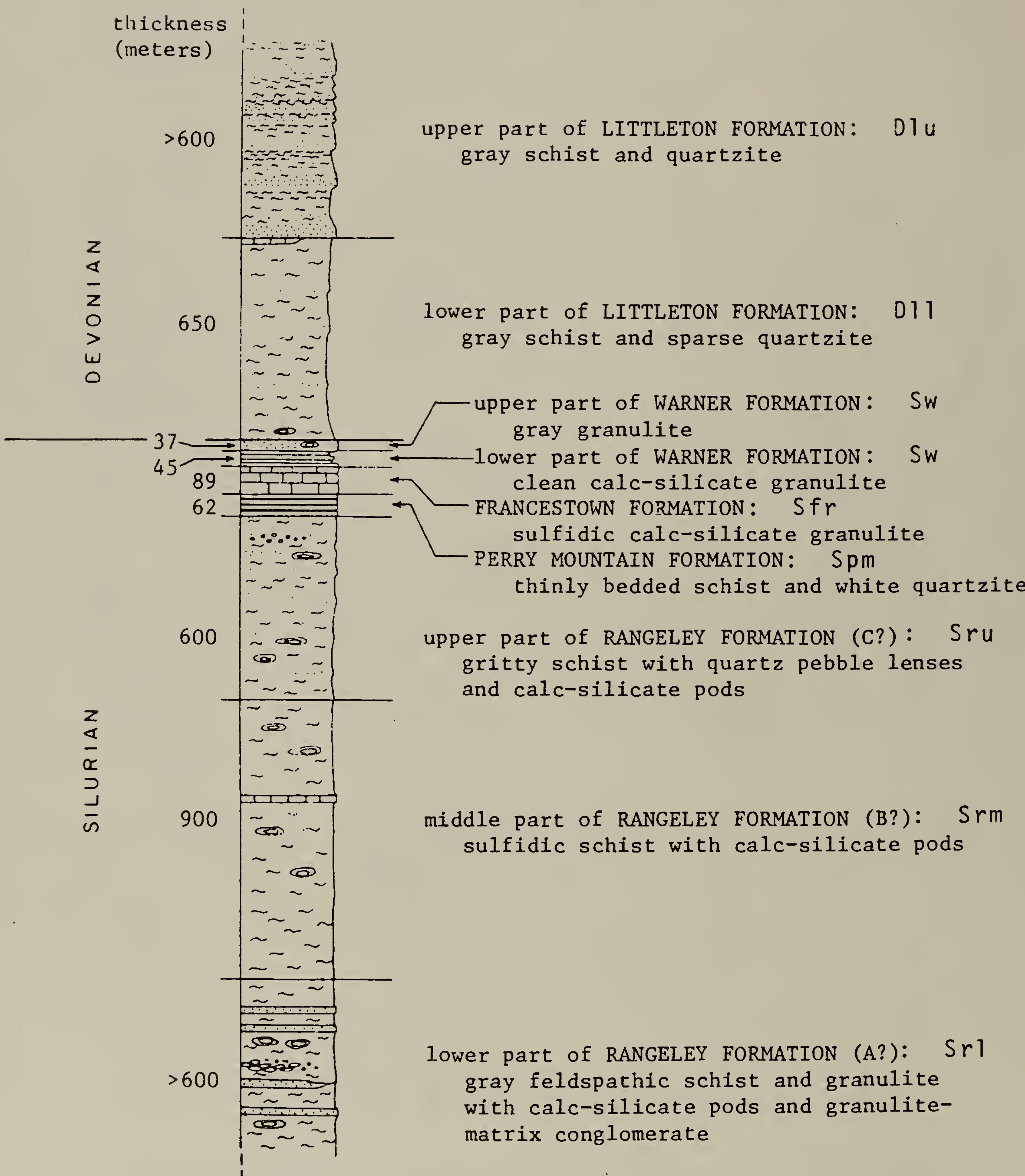



Fig. 2. Simplified geologic map of the Monadnock quadrangle. Trip B-2 stops are shown by circled numbers. Summit of Mt. Monadnock (Trip C-1) is a solid triangle. Numbers in parentheses refer to the three structural levels separated by faults discussed in the text. Littleton Fm.—dotted; plutons—shaded; Rangeley Fm. and other older units—unpatterned. (Littleton not shown in Derby Hill window: see Figs. 10 & 11).

Fig. 3. Stratigraphic Column.



schists with gritty horizons, calc-silicate pods, and local lenses of quartz-pebble conglomerate. Bedding is generally well defined in the upper part and beds up to 15 cm thick with "slow grading" are present.

Reddish-rusty, quartz-biotite-muscovite-plagioclase schist, with gritty horizons and calc-silicate pods, is the most common rock type in the Rangeley Formation. The plagioclase ranges from $An_{23}$ to $An_{55}$. Sillimanite, garnet, orthoclase, and retrograde staurolite and Mg-chlorite may be present. Pyrite, pyrrhotite, and marcasite after pyrrhotite account for the rusty weathering. Lenses or pods of calc-silicate granulite are very common, ranging in size from 10 to 100 cm. The discovery of conglomerate lenses at some fifteen localities lends weight to the proposed correlation of this unit with type Rangeley Formation in Maine.

Gray-weathering schists in the Rangeley are quite similar to the Littleton Formation. The Rangeley contains somewhat more feldspar and fewer phyllosilicate minerals. A more reliable characteristic of the Rangeley gray schists is the presence of calc-silicate pods similar to those in the rusty schists.

In some parts of the Rangeley Formation, but especially in the gray-weathering schist and gneiss, segregations of quartz and muscovite, with or without feldspar and sillimanite, form augen. Augen are especially common north of the Chesham Pond thrust. They may represent pockets of recrystallized pegmatitic melt or retrograded K-feldspar porphyroblasts. The augen textures may have been accentuated by mylonitization early in the deformation.

Perry Mountain Formation

In the Monadnock quadrangle an interval of thinly bedded (2-5 cm) clean quartzite and gray to slightly rusty schist occurs in most places between the Rangeley and Francestown Formations. Quartzite-schist contacts are sharp, and graded beds are sparse. Where graded beds are present the topping directions are commonly hard to read. Norm Hatch (pers. comm., 1985) thinks that much of what I mapped as Perry Mountain should be included with the upper Rangeley. Some outcrops in the upper part of the Rangeley Formation resemble the Perry Mountain, and indeed the contact may be gradational. East of the Thorndike Pond fault zone, little if any Perry Mountain Formation can be recognized at the appropriate position.

The quartzites consist of quartz, muscovite, biotite, garnet, and opaques, with or without sillimanite, plagioclase, retrograde chlorite, and accessory apatite, zircon, and tourmaline. The schists are gray to somewhat rusty-weathering, quartz-biotite-plagioclase-garnet rocks with or without sillimanite, muscovite and orthoclase. Locally there are large sillimanite pseudomorphs after andalusite. Thin garnet-quartz granulite ("coticule") beds and lenses are locally present in the Perry Mountain Formation.

The Perry Mountain Formation is much thicker in the Peterborough quadrangle (Duke, 1984) and correlates with the upper part of the Crotched Mountain Formation in the Hillsboro quadrangle (Nielson, 1981). The clean quartzites and thinner bedding compared to the underlying Rangeley Formation suggest a maturing of the erosional cycle in the source area. The Clough Quartzite may be in part correlative (Figure 1), but no fossils have yet been found in the Perry Mountain Formation.

Francestown Formation

Extremely rusty-weathering, blocky calc-silicate granulite and rusty-weathering graphitic schist make this the most distinctive unit in the Monadnock stratigraphy. Graphite and sulfides together compose up to 10% of the rock. The calc-silicate granulites are composed of quartz, calcic plagioclase, graphite and iron sulfide minerals with or without sphene, actinolite, diopside, zoisite, microcline, minor muscovite, Mg-biotite, and Mg-chlorite, and accessory apatite, zircon, tourmaline, and allanite. The iron sulfides include pyrite, pyrrhotite and secondary marcasite apparently replacing pyrrhotite. The rocks are hard and well jointed, breaking into brick-sized fragments.

Schists of the Francestown Formation weather deeply to a rusty brown-orange-yellow rind surrounding a white interior flecked with graphite. The schists contain quartz, muscovite, Mg-biotite, sillimanite, graphite, iron sulfides, and secondary Mg-chlorite, with or without plagioclase, rutile, sphene and zircon. Neither the schists nor the calc-silicates contain black mica, and this is one of the chief criteria for distinguishing isolated outcrops from otherwise similar rusty schists in the Rangeley Formation. The correlative Smalls Falls Formation (Nielson, 1981) contains a higher proportion of schist to calc-silicate than in the Monadnock area.

Warner Formation

Although they have not been mapped separately, two parts of the Warner Formation can be distinguished in the Monadnock quadrangle. The lower part of the Warner Formation consists of thinly bedded (0.5 - 5 cm) green, pink, gray or white calc-silicate granulites. The color variation is due to differences in the relative proportions of actinolite, diopside, garnet, clinzoisite, ferrian zoisite, sphene, biotite and calcite. Quartz is present in all layers. Rare schist beds occur near the base of the formation. The upper part of the Warner consists of fine-grained quartz-biotite-plagioclase granulite, with minor muscovite and sphene, and accessory zircon, garnet, apatite, rutile, tourmaline and opaques. The plagioclase is andesine. Outcrops are thickly bedded, massive, with smooth purplish-gray "salt and pepper" surfaces. Calc-silicate pods with mineralogy similar to the lower part of the Warner are common in the upper part. The pods are generally zoned with a weathered-out depression around a very resistant core and are similar to pods in the Rangeley Formation. The contact with the overlying Littleton Formation is gradational; calc-silicate pods and granulite beds persist into the lower part

of the Littleton. For mapping purposes, the uppermost continuous granulite bed is taken as the upper contact of the Warner. The Warner probably correlates with the late Silurian Fitch Formation of the Bronson Hill sequence, and with the Madrid Formation of Maine (Hatch et al., 1983).

Littleton Formation

The Littleton Formation in the Monadnock quadrangle consists predominantly of gray-weathering pelitic schist and micaceous quartzite. The schists consist of quartz and muscovite with or without biotite, staurolite, garnet, sillimanite, plagioclase, K-feldspar, graphite and other opaques, retrograde chlorite, chloritoid and staurolite, and accessory tourmaline, zircon and apatite. Pseudomorphs of sillimanite after andalusite are common in some parts of the quadrangle, depending on the metamorphic history, but no relict andalusite was found in thin section. Some pseudomorphs preserve chiastolite cross-shaped inclusion patterns.

The quartzites contain quartz, muscovite, biotite, garnet and opaques, with or without sillimanite, plagioclase, retrograde chlorite, and accessory tourmaline, zircon and apatite. The proportion of quartzite to schist increases in stratigraphically higher parts of the formation, so that two parts can be informally distinguished. Quartzite beds are progressively thicker and more abundant upward in the section. The upper part is best exposed on the summit of Mt. Monadnock, on Pumpelly Ridge, and around the village of Dublin. Coticule horizons are common in the upper part of the Littleton.

The Littleton of the Monadnock area is correlated, on the basis of both rock type and stratigraphic position, to the type Littleton Formation and to the Seboomook Formation of Maine, both of which contain Lower Devonian fossils.

## STRUCTURAL GEOLOGY

Introduction

Five phases of Acadian deformation have affected the rocks of the Monadnock quadrangle. The earliest, isoclinal folds, are believed to be related to the huge west-verging nappes proposed by Thompson et al. (1968). West-verging ductile thrust faults then developed and cut across the axial surfaces of the earlier fold nappes. The fold nappes and thrust faults are deformed by two phases of folds and backthrusts related to a complicated "backfolding" episode, and by folds related to the rise of gneiss domes. A summary of the structural history is presented in Table 1. Figure 4 shows the axial traces of the major folds and their relative ages.

The nappes and thrusts transported hot rocks onto relatively cooler rocks, setting up temperature and pressure gradients which led to peak metamorphic conditions closely following the nappe stage. The dominant foliation is parallel to axial planes of nappe-stage folds. Reactions proposed to explain garnet zoning

Table 1. Summary of structural history in the Monadnock quadrangle.

| Age | m.y. | Event | Description |
|---|---|---|---|
| Mesozoic | 245-144 | Extensional faulting; silicified zones; diabase dike. | |
| Late Paleozoic | 326-245 | Continued local shearing; ?Permian metamorphic disturbance? | |
| ?Mississippian | 326 | Intrusion of Fitzwilliam Granite plutons and microdiorite dikes. | |
| Devonian (Acadian orogeny) | | Late open folds | NW-trending, steep AP's; local crenulation cleavage. |
| | <380 | DOMING | Various trends, steep AP's; crenulation cleavage; strong linear fabric swirl |
| | | LATE BACKFOLDING | NE-trending fold axes, inclined AP's; linear fabric? |
| | | EARLY BACKFOLDING | W-over-E verging reclined folds; local mylonitization and mylonitic foliation; some linear fabric; peak of metamorphism? |
| | 393 | Intrusion of Spaulding Series plutons. | |
| | | THRUST NAPPES | E-over-W verging ductile thrust faults. |
| | | FOLD NAPPES | E-over-W verging isoclinal folds; pervasive foliation; quartz lineations? |
| | 413-402 | Intrusion of Kinsman Granite. | |
| Ordovician to Devonian | 480-415 | Deposition of volcanic and sedimentary rocks. | |
| ?Proterozoic Z to ?Ordovician | 600-480 | Genesis of rocks forming cores of later gneiss domes. | |

Summary of structural features in Monadnock quadrangle.

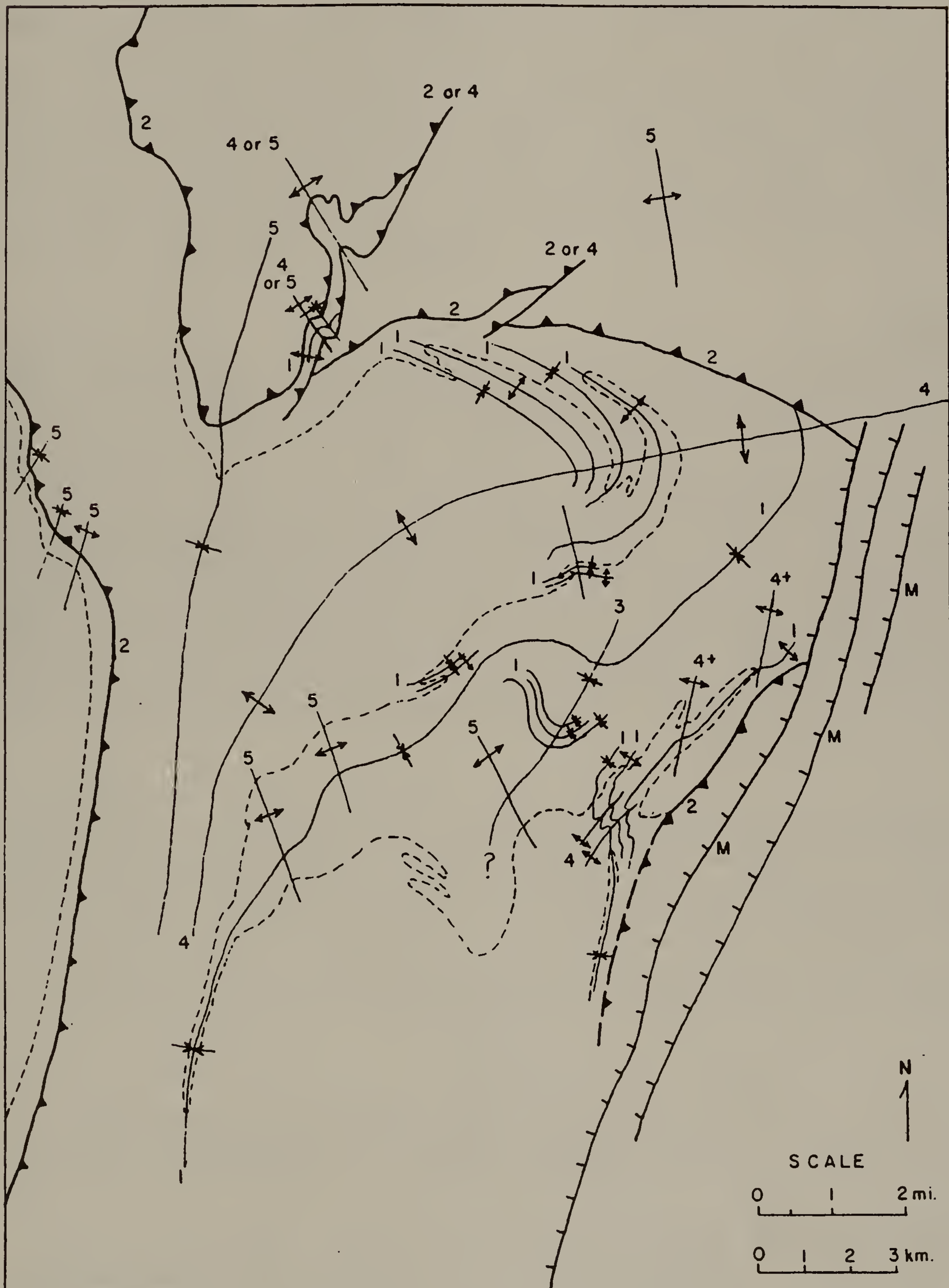



Fig. 4. Axial traces of major folds and faults, showing phases of deformation: (1) nappes, (2) thrusts, (3) early backfolds, (4) late backfolds, (5) dome-stage and other late folds, (M) Mesozoic faults. Base of the Littleton Formation (dashed line) is shown locally for reference. Names of structural features are shown in Figure 2.

(see metamorphism section) imply a declining pressure just beyond the peak of metamorphism as a result of uplift. However, thermal equilibrium was apparently not attained throughout the pile of nappes, resulting in an inverted metamorphic sequence, with lower grade rocks at the lowest structural level next to the Keene dome. The boundaries between Zone II and III metamorphic assemblages and between Zones III and IV (Figure 5) roughly follow the nappe-stage thrust faults, but there are no sharp discontinuities in metamorphic grade across the faults. A large backfold (Beech Hill anticline) deforms the Zone III-Zone IV boundary. The rocks were in the stability field of sillimanite during backfolding and doming.

Fold Nappes

The Monadnock syncline is interpreted as a nappe-stage fold. It is separated into two parts by intrusions (Figure 2). Southwest of the intrusions, the syncline is relative narrow, with good symmetry across its axial trace. North of the intrusion it is wider, and overturned to the southeast. Although the stratigraphy is grossly symmetrical across it, the presently upright southeast limb is much thicker than the presently overturned northwest limb. Prior to backfolding, these topping directions are believed to have been opposite to what they are today. The southeast limb includes a thick section of the upper part of the Littleton Formation, at least in part thickened tectonically. The southeast limb is folded by the younger northeast-trending Thoreau Bog syncline, but the northwest limb is not affected by it. This younger fold is apparently disharmonic, dying out rapidly to the north.

Nappe-stage folds are abundantly exposed on Mt. Monadnock. They consist of tight isoclinal folds with amplitudes far exceeding their wavelengths, and axial planes parallel to the pervasive foliation. Amplitudes are generally greater than five meters. The best exposed nappe-stage fold is the "Billings' fold" on a west-facing cliff 450 feet west of the summit. The cliff provides a view nearly perpendicular to the fold axis. It is a recumbent syncline. Layers are much thicker in the hinge than on the limbs.

Several map-scale folds on Figure 2 are interpreted as nappe-stage folds: Howe Reservoir syncline, Dublin Pond syncline, and Gilson Pond anticline. There are also isoclinal nappe-stage folds, deformed by younger structures, in the Derby Hill "window".

Early Thrust Faults

Three tectonic levels are present in the Monadnock quadrangle, apparently separated by early thrust faults (the Chesham Pond fault and the Brennan Hill fault, Figure 2). These levels contain: (1) an autochthonous, upright sequence overlying the Keene dome, consisting of Ammonoosuc Volcanics, Partridge Formation, Clough Quartzite, local Fitch Formation, and Littleton Formation, (2) the folded Monadnock sequence consisting of the units shown in the stratigraphic column of Figure 3, intruded by Spaul-

ding Tonalite, and (3) a "gneissified" sequence consisting mainly of Rangeley Formation and Kinsman Granite. Gravity studies show that the Cardigan pluton forms a 2-3 km thick, subhorizontal sheet-like mass (Nielson et al., 1976). Partly because the Kinsman and related Bethlehem Gneiss are exposed above the Bernardston and Skitchewaug nappes to the west, they are believed to have intruded prior to the formation of the nappes (Thompson et al., 1968). The Kinsman is apparently cut by the Chesham Pond fault. Evidence in the Monadnock quadrangle indicates that Spaulding Tonalite is younger than the nappes.

The Brennan Hill thrust fault is interpreted as a fault largely on a stratigraphic basis. The contact is difficult to map. To the west are sillimanite-rich, locally staurolite-bearing, monotonous gray-weathering schists of the Littleton Formation. To the east are more feldspathic gray schists of the Rangeley Formation, with abundant quartz-feldspar augen.

The Chesham Pond fault is much better documented than the Brennan Hill fault. North of the Spaulding pluton, Littleton Formation at the top of an upright sequence is overlain by augen schist of the Rangeley Formation. The contact is not exposed, but can be narrowed down to less than ten feet in a few places. To the east the Chesham Pond thrust is deformed by the Beech Hill anticline, and merges with the Thorndike Pond fault zone, where younger faulting obscures its exact location. The Monadnock syncline is cut by the Chesham Pond fault.

Backfolding and Backthrusting

There is a wide range of folds which deform the dominant foliation in the Monadnock quadrangle. In several outcrops nappe-stage isoclinal folds are deformed by younger folds, but there are very few instances of intersecting younger folds, so that little success has been made in sorting out the folds of intermediate age. Open folds with steep axial planes and associated crenulation cleavage are obviously younger and are related to the Keene dome. West-over-east asymmetric folds which deform the dominant foliation are very common on Mt. Monadnock, and have the same sense and style as the Thoreau Bog syncline (Figure 2 and Trip C-1, Figure 2).

The Beech Hill anticline is a major backfold which plunges moderately northwest. At first it appears to be paired with the Monadnock syncline because both are overturned toward the southeast. However, my present interpretation is that the Monadnock syncline is an older structure in the southeast limb of the Beech Hill anticline. This overturned limb dips steeply northwest, and continues south across central Massachusetts. It is presumably a short limb in the backfold system. Deep seismic reflection profiling to the north (Ando et al., 1984) suggests that the steeply dipping structures do not continue very far below the earth's surface. Spaulding Tonalite and Kinsman Granite were locally mylonitized during backfolding on the southeast limb of the Beech Hill anticline in the Thorndike Pond fault zone. This

zone was probably reactivated in the Mesozoic.

Backthrusts cut the Chesham Pond fault, most notably at the Derby Hill "window", where an area of folded rocks of the Monadnock sequence is surrounded by augen schists and rusty gneisses of the Rangeley Formation. At first I interpreted this area as a true window in the upper plate of the Chesham Pond fault, but a more reasonable explanation involves an east-directed backthrust (Trip B-2, Figures 5 and 6).

Doming

The Keene dome is one of the many gneiss domes along the Bronson Hill anticlinorium which, due to their relatively low density, rose late in the Acadian orogeny (Thompson, et al., 1968). The domes are characteristically separated by tight synclines in the mantling metamorphic rocks. There are abundant upright, east-over-west minor folds with associated crenulation cleavage up to three miles east of the Keene dome. North of the latitude where Rte. 12 crosses the edge of the dome, minor folds plunge gently north. To the south, most plunge south. Foliation and bedding next to the dome dip moderately east. Farther east dips steepen to vertical and beyond.

The rotation sense of dome-stage folds reverses across the Marlboro syncline, which is interpreted as a dome-stage structure which deforms the Chesham Pond thrust (Figure 2). The syncline strikes north toward an area of a high positive Bouguer gravity anomaly (Nielson et al., 1976) which is centered on the Rangeley Formation west of the Cardigan pluton.

Summary

Rocks of the Merrimack trough were deformed by fold-nappes and thrust-nappes directed toward the west, followed by backfolds and doming. The Brennan Hill fault transported rocks of the "Monadnock sequence" over the thinner, autochthonous sequence of the Bronson Hill anticlinorium. The thrust is interpreted here as a ductile thrust near the root zone of the Bernardston nappe. The Chesham Pond thrust carried the Kinsman Granite and Rangeley Formation westward over the Monadnock sequence, cutting across the nappe-stage Monadnock syncline between the Fall Mountain nappe and lower nappes. A major backfold, the Beech Hill anticline, deformed the Chesham Pond thrust, and it is in this anticline that the nappe-stage syncline is exposed. A similar interpretation can be extended northward. The "Kearsarge-Central Maine synclinorium" (Lyons et al., 1982), along which the New Hampshire sequence is exposed beneath a sheet of Kinsman Granite, is probably the same nappe-stage syncline exposed because of a younger backfold anticline. According to this model, the Fall Mountain nappe and the Kinsman Granite must be rooted east of the Kearsarge-Central Maine synclinorium. Inverted units of the Monadnock sequence appear locally beneath the west edge of the Ashuelot pluton of Kinsman Granite southwest of Keene (Elbert, 1986).

## METAMORPHISM

### Metamorphic Assemblages in Pelitic Schists

In the Monadnock quadrangle regional Acadian metamorphism has affected all the layered rocks and the Kinsman Granite. The peak of metamorphism produced predominantly "upper sillimanite" zone mineral assemblages in the pelitic schists: Zone III (sil-mus-gar-biot) and Zone IV (sil-mus-gar-biot-ksp) of Tracy (1975). The Chesham Pond fault coincides approximately with the boundary between Zones III and IV. Rocks containing staurolite (Zone II: sil-mus-st-gar-biot) occur west of the Brennan Hill fault. Cordierite was observed in thin sections of rocks immediately west of the Cardigan pluton (Zone VI: sil-gar-biot-ksp-crd). Sillimanite pseudomorphs after andalusite("andalumps") are abundant in much of the quadrangle. A line on Figure 5 shows the approximate former position of the triple point between the three aluminum silicate polymorphs, kyanite, andalusite, and sillimanite. Kyanite was transformed to sillimanite west of the line, and andalusite was replaced by sillimanite to the east. Neither relict kyanite nor relict andalusite has been found in the quadrangle.

Retrograde replacement of sillimanite by muscovite and of K-feldspar by muscovite and quartz is widespread in the quadrangle. More advanced retrogression occurred in the area northeast of Mt. Monadnock, where Fe-chlorite has partially replaced garnet, and secondary muscovite and chlorite are major components of the rock. Chloritoid is also common on Mt. Monadnock and in the retrograded area. In hand sample, it appears as dark green clumps which weather out to form pits. Tiny staurolites were observed in thin section with chloritoid, chlorite, and muscovite, replacing sillimanite and biotite. Brownish tourmaline in hand samples can easily be mistaken for staurolite, but prograde staurolite was not found anywhere outside of Zone II shown on Figure 5.

### Garnet Zoning in Zone III and P-T Estimates

Two garnets from a typical Zone III pelitic schist (MK-432) were studied with the electron microprobe. Pyrope increases toward the rim, except where the garnet touches biotite, and there pyrope decreases abruptly. Almandine is fairly constant except toward rims touching biotite, where it rises. Spessartine follows a pattern inverse to that of pyrope. Grossular shows little variation. One possible explanation for these trends is early prograde garnet growth from staurolite + biotite + quartz. After staurolite was consumed, garnet growth stopped and was later affected by local retrograde cation exchange where garnet touched muscovite and biotite. Two biotite-free garnet rim analyses compared with matrix biotite give temperature estimates of 670 and 635° C, using Thompson's (1976) calibrations. Spear and Selverstone's (1984) garnet and biotite isopleth diagram yields an intersection at 650° C and 5.8 kbar for the same data.

Garnet zoning in a different sample (MK-629) with the unusual K-poor assemblage quartz-biotite-plagioclase-garnet-cordierite-sillimanite, shows a decrease in pyrope and spessartine

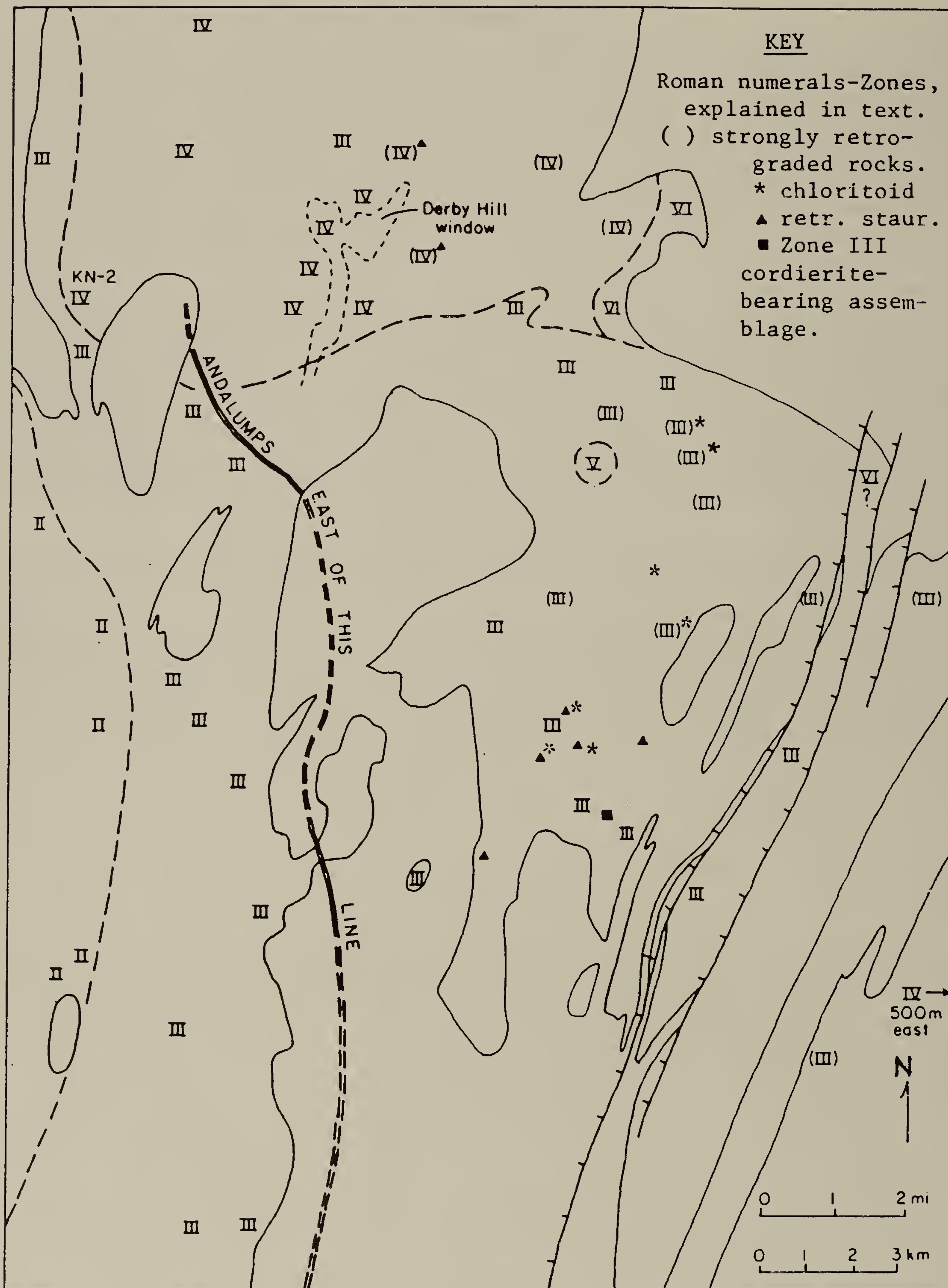



Fig. 5. Metamorphic zones based on assemblages observed in thin section, or staurolite in hand sample (Zone II). Includes data from Chamberlain (1981). Plutons outlined for reference.

toward the rim, and an increase in almandine. The most magnesian garnet composition and a temperture of 640° C give an estimated pressure of 6.3 kbar using Tracy et al.'s (1976) calibration of the reaction [sil + gar + qtz = crd]. The garnet rim composition in contact with cordierite, and the temperature estimated from core cordierite composition (620° C) gives a pressure of 6.1 kbar. These estimates seem consistent with the concept that the garnet zoning in MK-629 took place under conditions of falling temperature, but the zoning trend is not well understood.

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## Trip B-2: Stratigraphy and structure along the Chesham Pond thrust and Thorndike Pond fault zone, Monadnock quadrangle

### ITINERARY

Assembly point is Keene State College, Commons Parking lot, 8:30 a.m. Topographic Maps: Monadnock 15' quadrangle, OR Marlborough and Monadnock Mountain 7.5' x 15' provisional edition (1984) quadrangles.

Permission has been obtained from private landowners for the 1988 field trip; if you wish to visit Stops 4, 8, 10, 11A, 12, 13 on your own, please secure landowners' permission.

| Mileage | |
|---|---|
| 0.0 | Take Wyman Way to Main Street. |
| 0.2 | Right (S) on Main Street. |
| 0.5 | Straight through lights on Rte. 12 south. |
| 4.5-4.6 | Roadcut on left: east edge of Keene dome, overlain by Ammonoosuc Volcanics, Partridge Fm., Clough Quartzite, and Littleton Fm. |
| 4.9 | Approximate location of Brennan Hill thrust. |
| 5.3-5.4 | Roadcut on right: gray Rangeley Fm. |
| 5.6-5.7 | STOP 1A: Rangeley Formation. (Quick stop to observe lithology.) Roadcuts on both sides: rusty quartz-biotite-plagioclase-sillimanite-muscovite schist with beds and pods of calc-silicate granulite. Layering and foliation dip moderately east. Mineral lineations and crinkle fold axes (dome stage) plunge 35 to 45° SE. Boudinaged pegmatite at east end of cut.<br><br>Continue east on Rte. 12. |
| 6.2 | STOP 1B: Rangeley Formation. Park in rest area on right.<br><br>Road cut on south side: gray augen schist, with minor calc-silicate pods. East-over-west dome-stage folds plunge south. Any specula- |

tion as to the origin of the augen?
Roadcut on north side: gray Rangeley with rusty Rangeley at east end.

Continue east on Rte. 12.

6.4-7.0 Rusty Rangeley continues.

7.5 Rusty Rangeley roadcut on left beyond bridge.

8.2 Cemetery on left. Crossing Francestown and Warner Fms. on west limb of Monadnock syncline (Fig. 6).

8.9 Troy village green.

9.0 At south end of green, turn right (W) on High Street, past bakery.

9.05 Cross old RR bridge -- RR cut is in Littleton Fm.

9.1-9.2 Littleton outcrops in house yards (return here for Stop 3).

9.4 Recrossing Francestown Fm., west away from Monadnock syncline.

9.6 Turn right up dirt driveway (across from white garage with red doors). Rangeley outcrops in woods to east: rusty biotite-sillimanite schist.

9.7 Turn vehicles around at end of drive.

STOP 2: Quartz-pebble conglomerate in Rangeley Formation.
Walk down drive, look for stonewall in woods to left (E). Outcrop is 50 feet south of wall. Contact between rusty micaceous quartzite and conglomerate strikes N04E, dips 56 NW. Pebbles are flattened

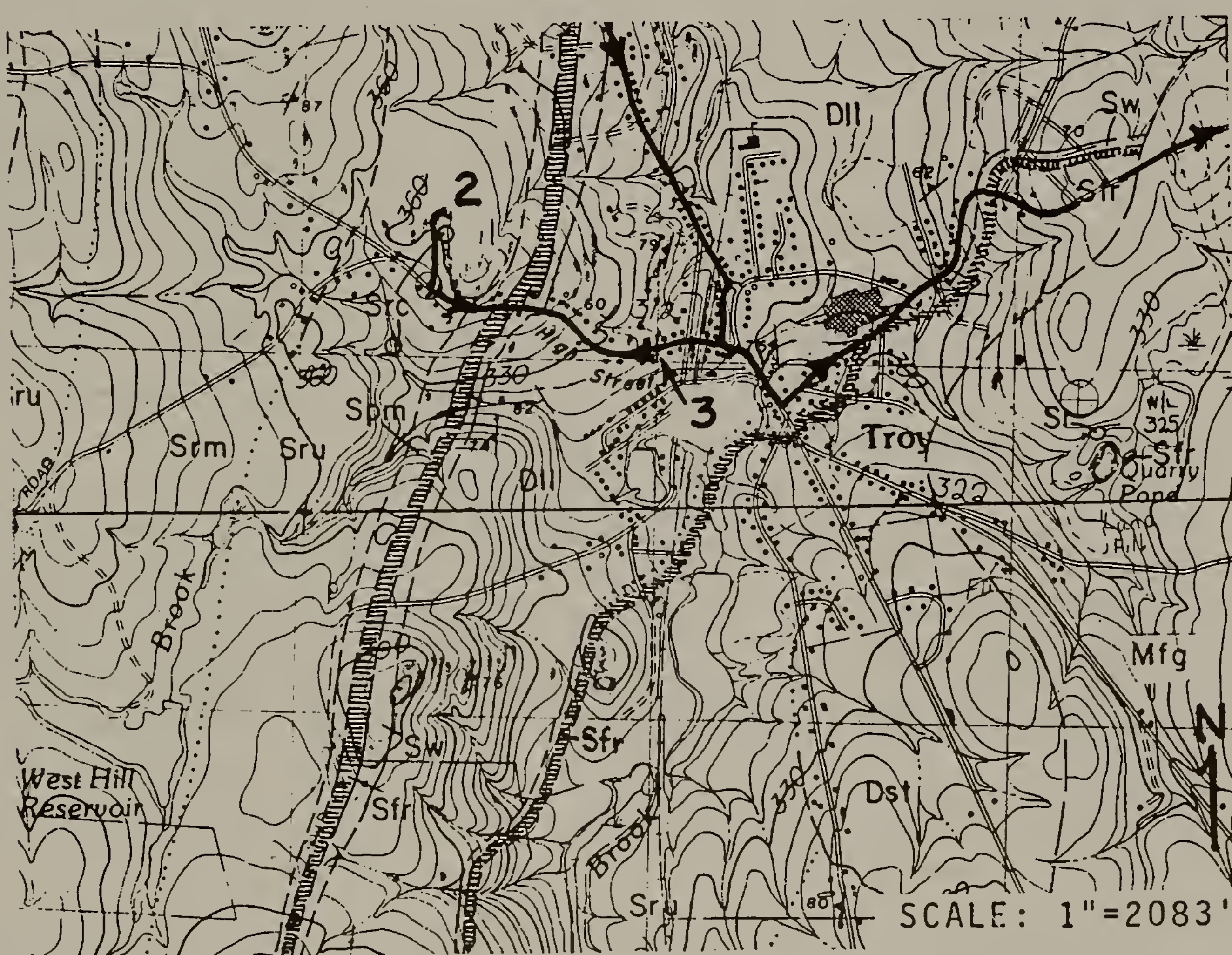


Fig. 6. Monadnock syncline in the Troy area, showing locations for Stops 2 and 3. Francestown ruled.

parallel to foliation oriented N14E, 66NW. Pebbles are vein quartz, white quartzite, and rare calc-silicate clasts which give the outcrop a pitted surface where they have weathered out. Although it is a grain-supported conglomerate, there is a calc-silicate matrix made of diopside, garnet, clinozoisite, actinolite, sphene and calcic plagioclase.

Retrace route toward center of Troy.

10.3 STOP 3: Littleton Formation. Small blasted outcrop to right (S) obscured by saplings. Spangles of muscovite across the foliation are typical of Littleton in the southern part of Monadnock syncline. We are west of the aluminum silicate triple-point isobar.

10.4 Troy green. Go briefly east on Rte. 12 toward Fitzwilliam.

10.5 Pond on right

10.6 Turn left on road to Jaffrey and Mt. Monadnock.

11.2 Crossing Warner and Francestown on east limb of Monadnock syncline.

11.7 Stonewalls and large boulders of Fitzwilliam Granite.

12.3 View to east of Gap Mtn.: Littleton with sillimanite pseudomorphs after andalusite on summit, in roof pendants within Spaulding Tonalite.

13.2 Turn right (E) on Rte. 124.

13.9 Spaulding Tonalite on right.

14.0 Littleton Fm. intruded by granite on left.

14.5 Parking area for White Arrow Trail to Monadnock (Trip C-1).

15.2 View north to Monadnock: Visible outcrops are entirely Littleton Fm., with Silurian units in woods along base on east limb of Monadnock syncline, and Spaulding Tonalite in the valley.

15.4-15.5 Spaulding Tonalite, both sides of road.

15.8 Glimpse of Cummings Pond on right; Rangeley Fm. outcrop on right.

15.9 Perry Mountain Fm. on right. Park on right side on wide shoulder near "School Bus Stop Ahead" sign:

STOP 4: Perry Mountain/Francestown contact. Don't disturb horses! Don't leave gates open! Walk down through horse pasture east of white house north of road to Dole Brook and contact between Perry Mountain Fm. (W) and Francestown Fm. (E). The contact is folded by open folds plunging 69° NE. Francestown here is mainly blocky-weathering granulite. Perry Mountain contains pitted horizons due to weathered-out biotite-garnet-apatite lenses, which may have been fossils.

Continue east on Rte. 124.

16.7 Obscure outcrops, both sides of road are Rangeley Fm.

16.9 Small outcrop on left: Kinsman Granite.

16.95 Jaffrey Center Fire Station. Thorndike Pond fault zone extends south through this area.

17.5 Left (N) on Dublin Road toward Monadnock State Park.

18.05 Small pond on right.

18.1 Recrossing layer of Kinsman that we passed by the fire station.

18.7 Pass Bible Conference Center, and turn left toward State Park, climbing up over a small drumlin.

19.1 STOP 5: Half-hour traverse across Gilson Pond anticline (Fig. 7). Park in small turnout on right with "No Picnicking" sign and follow trail to wildflower garden.


Fig. 7. Geology Southeast of Mt. Monadnock, showing Stop 4 and Stop 5, traverse along Ark Brook across an isoclinal fold. Horizontal ruling is the Francestown Fm.

STOP 5A: NO HAMMERS. DO NOT DISTURB VEGETATION OR PLANT LABELS. "The Boulder" in the garden is actually an outcrop of Littleton Fm.: biotite-garnet-sillimanite-muscovite-quartz schist, with thumbnail-sized garnets and some quartzose beds (graded beds top east). Smooth, dark gray calc-silicate granulite pods toward west side of outcrop are similar to Warner Fm., and are common in the Littleton up to 100 feet from the contact with Warner. A similar pod at Gilson Pond, one mile north, contains a core of quartz, diopside, calcite, bustamite, grossular and zoisite, grading into a rim of quartz, plagioclase, actinolite, clinozoisite, grossular and sphene next to the enclosing schist. At Stop 5A note the clumps of sillimanite up to 2 cm long. The Zone III metamorphic assemblage here has not suffered retrogression. Foliation is oriented N11E, 58NW.

From here we will go northwest across the Gilson Pond anticline which has Perry Mountain Fm. at its center. Follow trail east to stonewall and north along wall, through rhododendrons ("the Dell") to Ark Brook. Follow old Hinckley Trail up west bank. Low mossy outcrops in brook are Spaulding Tonalite.

STOP 5B: Strongly foliated "Gilson Pond type" Spaulding Tonalite in contact with Littleton Fm. to northwest. The tonalite is a biotite-plagioclase gneiss with flaser texture produced by two intersecting foliations. Feldspar phenocrysts in the Gilson Pond type are typically 1-2 cm long, and show a preferred orientation parallel to one of the foliations. Tails on sheared feldspars indicate a west-side-up sense of shear related to backfolding.

STOP 5C: About 150 feet upstream the brook branches. Warner Fm., intruded by Spauling Tonalite, is exposed in both branches. The typical slabby weathering of the Warner is best seen in the east (right) branch. Francestown Fm. crops out where the east branch rejoins the main brook. Both rusty, blocky-weathering calc-silicate granulite, and rusty, graphite-bearing, white mica schist of the Francestown are exposed. Foliation is oriented N24E, 74NW. Some surfaces have N-plunging lineations.

Continue up brook or trail about 300 feet.

STOP 5D: Perry Mountain Fm. along right bank; Francestown in stream bed west of the isoclinal axial trace. (Perry Mountain is better exposed at the next stop, on strike to the northeast.)

Continue upstream about 250 feet, passing Francestown outcrops in brook, to culvert and junction with Harling Trail which here follows an old road. Go northeast on road (XC 18): At 80' garnet schist, a discontinuous horizon at base of Warner Fm.; 120' to Francestown in road; 50' further to outcrops in woods to right of road:

STOP 5E: Perry Mountain Fm.: 3 to 4 cm interbedded white quartzite and schist, in the center of the isocline. About 800' NE of here the Perry Mountain hinges out, and Francestown occupies the center of the isocline for another 4000' where it is cut by the Spaulding.

Retrace route to Ark Brook at trail junction. Examine outcrops upstream from culvert.

STOP 5F: Warner Fm.: Banded calc-silicates of lower part of Warner can be seen in pool, with more massive feldspathic granulites in mossy outcrops beyond. 250 feet upstream from culvert, on right bank is a fold hinge plunging 42° to the north. This is an open backfold, west-over-east, of the sort that elsewhere deforms Spaulding Tonalite.

Warner Fm. crops out for another 850 feet upstream. Beyond that, Littleton Fm. continues to the top of Monadnock.

Retrace route to vehicles. CAUTION: avoid ski trail which leaves the brook; stay along the brook.

In vehicles, retrace route to Dublin Road.

19.5 Turn left (N) at stop sign onto Dublin Road.

20.7 Littleton Fm. on left.

20.9 Gilson Pond entrance to Monadnock State Park on left.

21.2 Thorndike Pond to right.

21.35 Turn right on dirt road at Camp Wanocksett. Spaulding Tonalite outcrops.

21.8 Crossing south end of a drumlin.

22.1 Keep straight (E) at intersection.

22.4 Bear left (dead end to right).

22.85 Turn right (E) on Craig Road (paved).

23.4 Stay on Craig Road, bearing left.

23.7 Bear left (N) on Rte. 137. Spaulding Tonalite west of junction.

24.0-24.1 Outcrops and boulders of Kinsman Granite on left.

25.6 Turn right (E) on Rte. 101 (Bonds Corner crossroads).

26.0 Pass stop 7 on left.

27.1-27.2 STOP 6: Roadcuts of Rangeley Formation at east edge of Monadnock quadrangle. Mainly a quick lithology stop: rusty schist with calc-silicate pods, much like Stop 1. Foliation and bedding dip steeply west. Many slickenside surfaces (Mesozoic?).

TURN AROUND WITH CARE! Proceed west on Rte. 101.

28.4 STOP 7: Outcrops of Kinsman Granite, north side. Foliation dips gently. Note folded mylonitized zone; xenoliths (calc-cilicates at west end); dark-colored rock at west end--is it mylonite or a xenolith?

28.8 At Bonds Corner stay straight west on Rte. 101.

29.8 View of Beech Hill straight ahead: Littleton Fm. in hinge of Beech Hill anticline (backfold stage).

30.3 Pass Monument Road on right, Hedge House Gifts on left. Park carefully on right shoulder before curb begins.

STOP 8: Upper part of Littleton Formation. Outcrop in trees south of Rte. 101, west of Hidden Brooks B & B. Glacially striated roches moutonnées of thinly bedded schist and quartzose schist with weathered-out retrograded sillimanite pseudomorphs and garnets. Best exposure south of stonewall. Similar to Littleton on Mt. Monadnock above the Seven Quartzites (see Trip C-1).

30.8 Blinking light in center of Dublin. Home of Yankee magazine.

31.0 Bear right on Old Common Road past Dublin School. Stay on this road to where it rejoins Rte. 101 at Dublin Pond, and park along it before stop sign. Do not park on Rte. 101.

31.5 STOP 9: Overturned west limb of Monadnock syncline (Fig. 8). WATCH TRAFFIC. Littleton Fm. in woods north of Old Common Road continues up to Beech Hill. Warner Fm. on sharp curve concave toward pond; Francestown on curve convex toward pond west of Old Harrisville Road. Spaulding-related granite intrudes the Francestown to the west. WATCH TRAFFIC ON BLIND CORNER!
Layering dips predominantly west. East-dipping Francestown is part of a minor fold. Minor folds in Warner are backfolds with east-over-west sense.

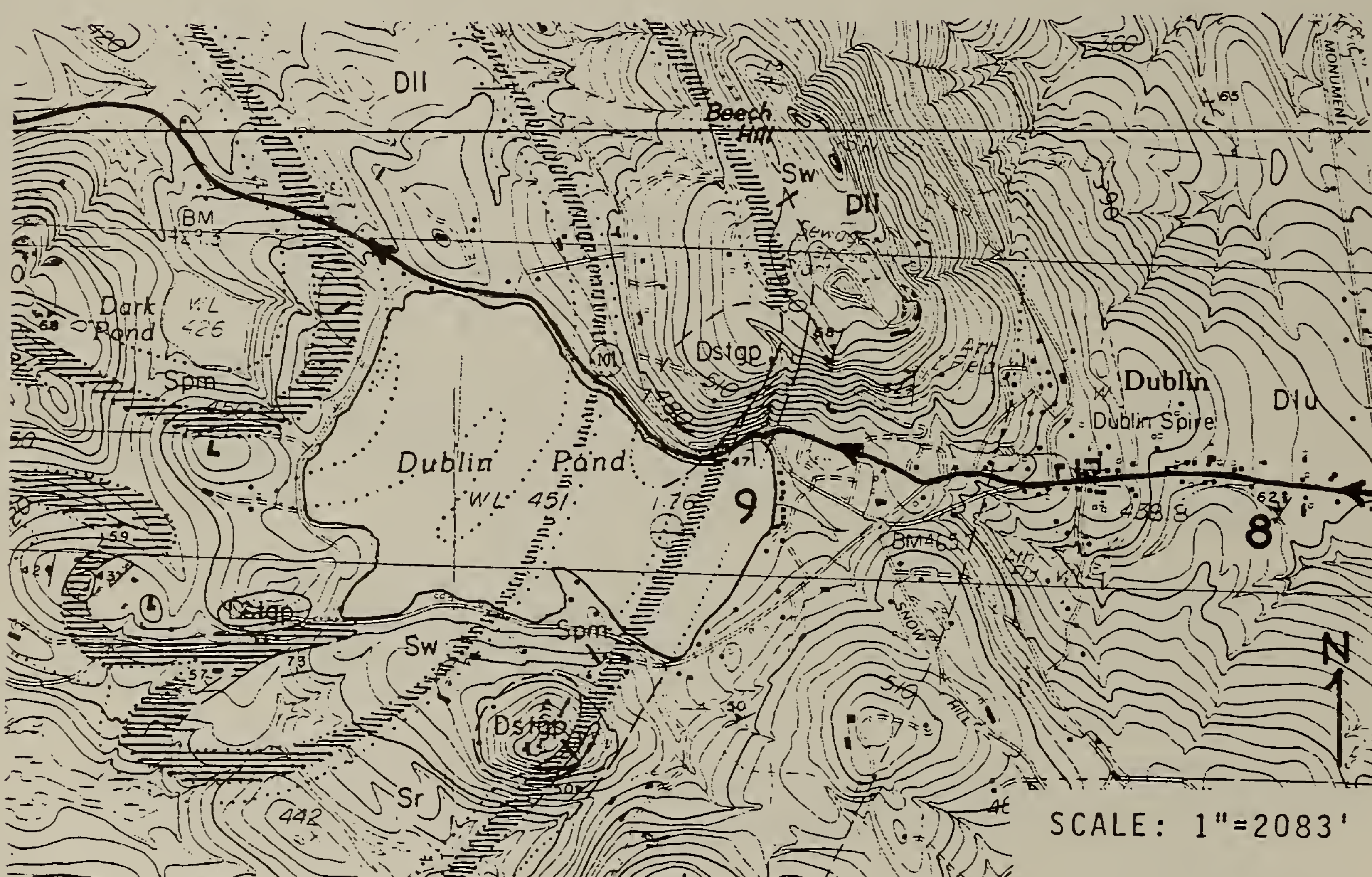


Fig. 8. Geology around Dublin Pond and Beech Hill (Stops 8 & 9). Horizontal ruling is the Francestown Fm.

Return to vehicles and continue west on Rte. 101.

33.4 Webber Nursery.

33.5 Rangeley Fm. on left.

34.3 Howe Reservoir. Spaulding Tonalite on left. (Type locality Spaulding Hill one mile to southwest.)

36.1 Marlborough town line.

36.3 Turn right (N) on Oliver Road.

36.6 At T intersection turn left onto Old Chesham Road. Stop 10 will visit an upright sequence which trends ENE parallel to Old Chesham Road ridge (Fig. 9).

36.7 STOP 10A: Outcrops in yard of log house. Watch out for rock garden plants. Perry Mountain Fm. with open folds and coticule.

Continue west on Old Chesham Road.

37.5 STOP 10B: Second old house on right beyond two drumlin hills. Proceed on foot to outcrops of Francestown Fm. in pasture on slope directly behind farm buildings. There are minor folds in the Francestown, but general trend is ENE.

Find gap in stonewall NW of outcrops, follow logging road into woods. Flagged route from road NW to outcrops of Warner and Littleton Fm. Gradational Warner/Littleton contact. Coarse (up to 1 cm) sillimanite is typical of Littleton in this area.

Return to vehicles and retrace route past Stop 10A, east on Old Chesham Road.

38.4 Keep straight east.

39.0 Bear straight (E) onto paved Chesham Road.

39.2 Cross bridge and turn left (N) on unmarked dirt road.

39.3 Francestown outcrops in woods to right, forming small ridge.

39.5 Stay straight on Old Harrisville Road (unmarked).

39.55 STOP 11A: Outcrops north of first house on right, in yard of guest house and directly north of guest house: Littleton Fm. with garnet and coarse sillimanite. Graded bed upside down. Calc-silicate pod. (Does this indicate proximity to Warner Fm.?)

Continue in vehicles northwest on Old Harrisville Road.

39.75 STOP 11B: Chesham Pond fault.
Pull onto shoulder beyond next house. Outcrops along roadbank: 100' to 200' beyond driveway: Littleton Fm.; 200' to 305': gap in outcrop; 305' to 325': Littleton Fm.--sillimanite schist; 325' to 400': Rangeley Fm.--massive augen schist.
Contact is not exposed, but there is no room for intervening units.

Continue west on Old Harrisville Road.


Fig. 9. Northwest limb of the Beech Hill anticline, showing Stops 10A and 10B on the upright sequence, Stop 11 at the Chesham Pond fault, and Stop 12 in the "window". Horizontal ruling indicates the Francestown Fm.

40.1 Turn right (N) on Hardy Hill Road. Rangeley Fm. in woods to right.

40.5 STOP 12: Traverse across southern part of Derby Hill "window". Rangeley Fm., mainly augen schist (metamorphosed mylonite?), surrounds an irregular area of younger rocks (Fig. 10). Within the area, Rangeley, Perry Mountain, Francestwon, Warner, and Littleton Fms. are deformed by isoclinal folds, which are in turn deformed by open folds. The window-like structure is interpreted as a backthrust along the east boundary of the window which displaces the Chesham Pond thrust (along the west boundary). We will follow a flagged route which zigzags across the structure.

Return to vehicles and continue north on dirt road to turn around in logging road at 40.8 miles if dry, or continue to house on hill to turn around.

42.05 Retrace route to Stop 11A.

42.1 Turn left (NE) on unmarked road.

42.7 Chesham Depot. Turn hard left toward Nelson.

43.0 Turn left on Sunset Hill Road.

43.5 Outcrops and boulders of Rangeley Fm. on left.

43.75 Keep straight on dirt road.

44.05 Road goes down over ledge of Rangeley Fm. Two driveways to right. Backthrust crosses road between here and Francestown outcrop on left (Figure 11).

44.40 Turn right on "Derby Hill Private Road".

44.45 Ask permission from Drury's before proceeding further.

44.55 Park near chimneys of burned-out building.
STOP 13: Derby Hill. Visit outcrops as time permits. Francestown outcrops at the chimneys were mapped by Fowler-Billings (1949) as the rusty quartzite member of the Littleton. Good exposures of Perry Mountain and Rangeley to west before reaching augen schist at scout camp. To the east, Warner is in contact with the augen schist near private road crossroads.

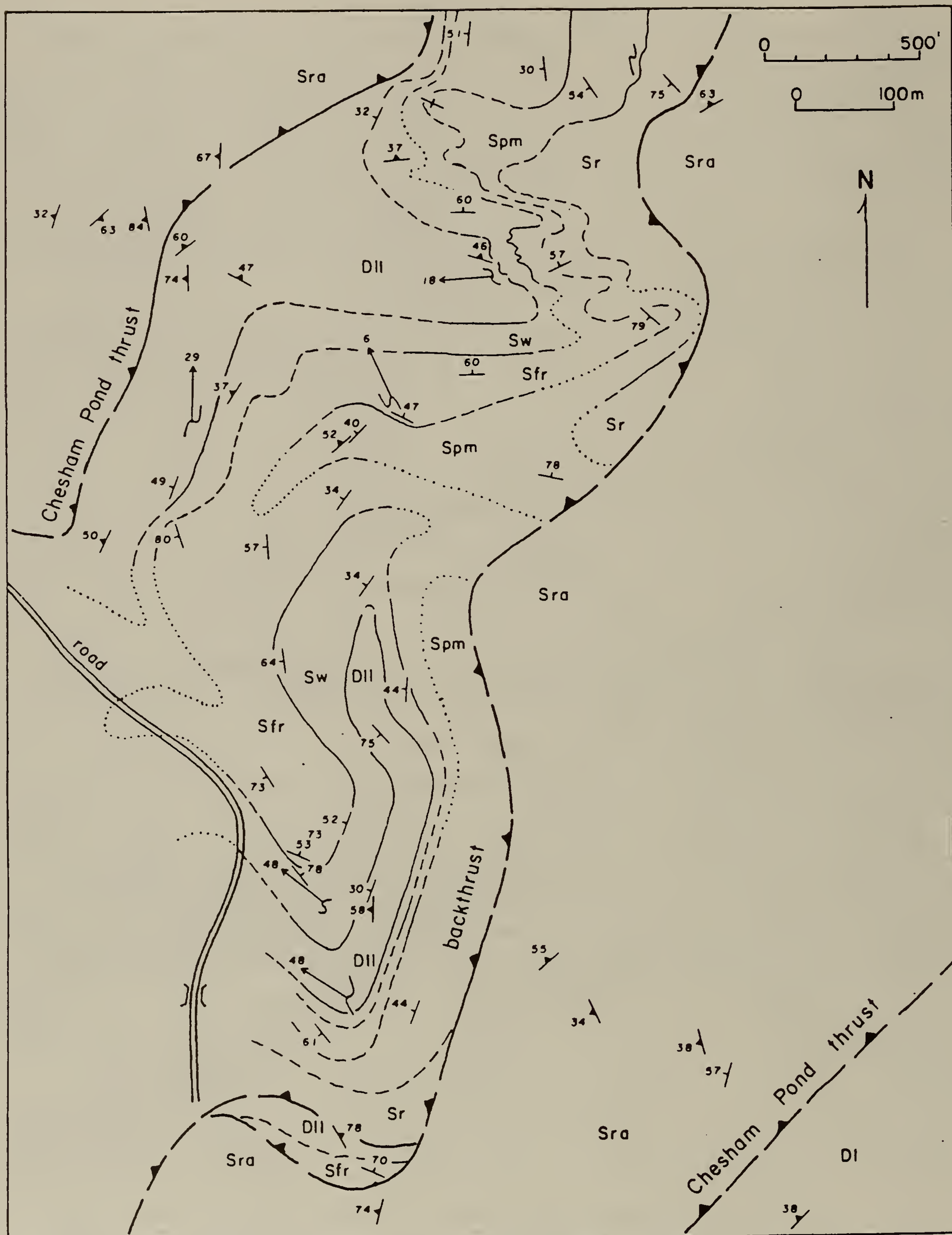


Fig. 10. South part of Derby Hill "window". Formations are abbreviated as on text Fig. 3. Sra is augen schist assigned to the Rangeley Fm., which in part may be metamorphosed mylonite. Stop 12 will cross both the backthrust and the Chesham Pond thrust.

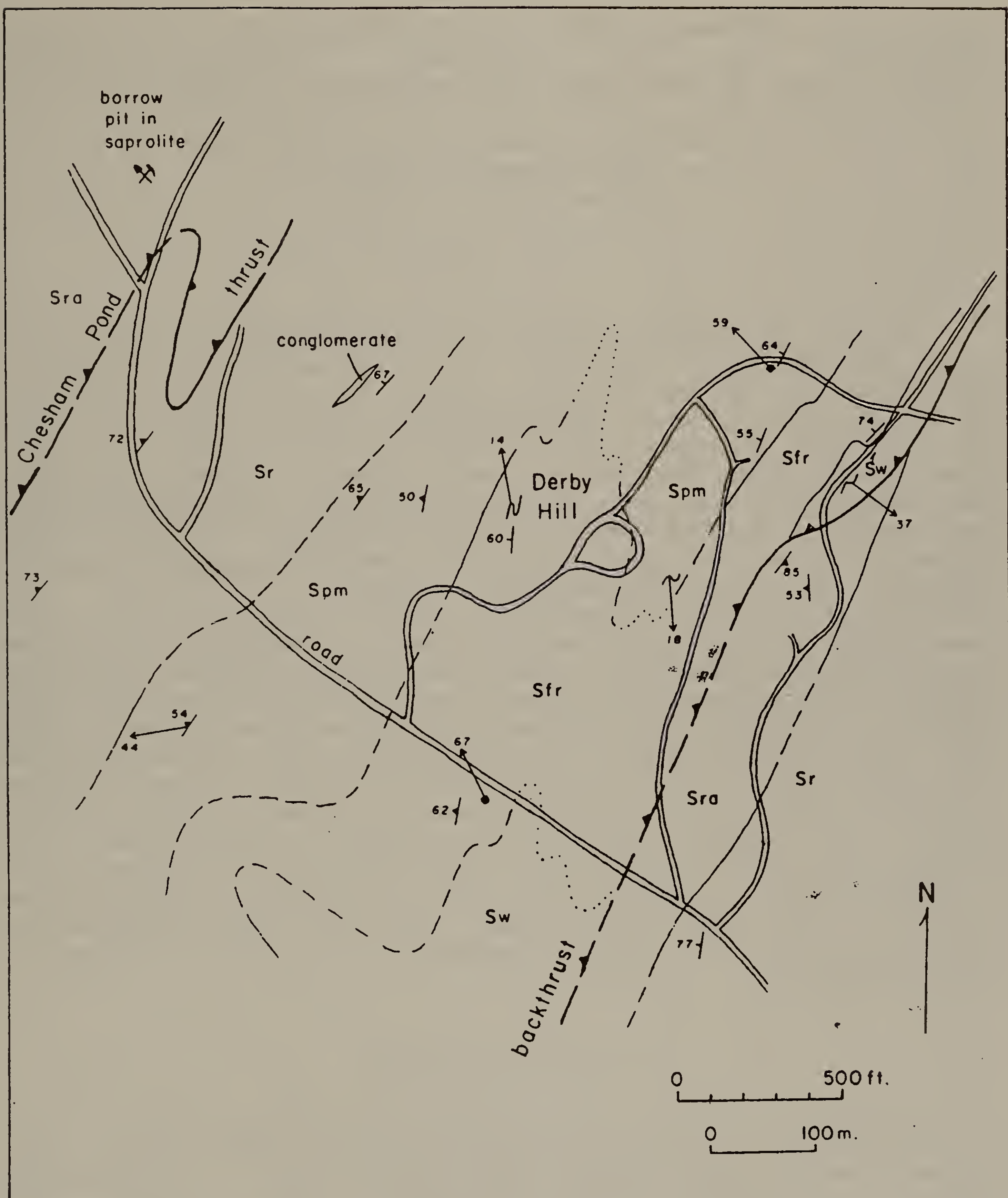


Fig. 11. A portion of the north part of Derby Hill "window". Abbreviations as on Fig. 3.and 10. Stop 13 will visit the Francestown Fm. on Derby Hill itself, and the backthrust along the driveways to the east.

# GEOLOGY OF THE MILLERS RIVER DELTA

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The Millers River Delta (Jahns and Willard, 1942) is a large outwash delta developed along the edges of pro-glacial Lake Hitchcock during late Wisconsinan ice-retreat. The delta has traditionally been attributed to sediment input from the Millers River (Emerson, 1898; Jahns, 1967; Brigham-Grette and Wise, 1988), but morphologic and geologic evidence suggest that most of the sediment was derived from glacial ice within the Connecticut River Valley. We, therefore, interpret this delta as a classic lacustrine ice-contact morphosequence (Koteff, 1974; Koteff and Pessl, 1981).

Lying at the junction between the narrow upper Connecticut River Valley and the broad southern Connecticut Valley, the Millers River Delta records the change from broad valley to narrow valley deglaciation during the retreat. The history of lake level lowering in northern Lake Hitchcock (the lake north of the delta which originally was separated from the main body of Lake Hitchcock by a short river flowing across the top of the delta) is recorded very nicely in this area. The field trip will examine evidence for the syndepositional position of the ice margin, discuss the probable source of the sediment, and examine evidence for the stages of the lake level drop. We will also discuss the probable effects of paleo-groundwater flow on the stages of river incision.

## GEOLOGIC SETTING

The Millers Falls Delta lies in the northeastern corner of the Mesozoic graben that creates the broad Connecticut Valley through Massachusetts and Connecticut. To the north the Connecticut River flows through a narrow valley that roughly follows the boundary between the phyllites of the Connecticut Valley Synclinorium and the gneisses of the Eastern Highlands. The delta occupies the junction between the broad and narrow portions of the Connecticut Valley. The Millers River Valley is incised into the Paleozoic gneisses of the Eastern Highlands and joins the Connecticut Valley within the area choked by the delta.

During ice retreat in central New England, a large pro-glacial lake (Lake Hitchcock) filled the ice free portions of the Connecticut Valley (for a summary of the late glacial history of the Connecticut Valley, see Koteff and

others, 1988). The southern end of the lake was constrained by a dam created by a readvance moraine near Middletown, Connecticut, and the northern end approximated the ice margin in the valley as the ice retreated up the valley. The maximum extent of Lake Hitchcock was from about Middletown, Connecticut, to Peacham, Vermont. The demise of the lake occurred when the morainal dam was breached, draining the lake relatively rapidly. Because the Millers Falls delta divided the lake into a northern narrow portion and a southern broad portion, it had to be breached before the northern portion could be drained. Thus, the timing of lake level drop north of the delta post-dated that to the south.

## ICE RETREAT

Ice retreat in New England, including the Connecticut River Valley, is generally thought to have occurred by stagnation zone retreat (Koteff and Pessl, 1981). Basically, this mode of ice retreat entails a stepped retreat in which major sections of glacial ice stagnate, probably due to topographic impediments, and develop outwash features along their leading edges. As the stagnated ice melts, it gives way to normal valley (e.g. lacustrine) conditions and there is a new frontal edge to the ice sheet. It is hypothesized that the sediment for the outwash deposits is supplied by the ablation of the upthrusting edge of the active glacier (the dirt machine of Koteff, 1974).

In Connecticut Valley south of the Millers River delta, the nearest major outwash deposits are the Long Plain delta in Sunderland (Jahns, 1951) and the Barnes delta in Southampton and Westfield (Larsen, 1972). The Long Plain delta clearly developed after the basin to the south was clear of ice and is thought to record the presence of ice in the Montague basin north of Mount Toby (Brigham-Grette and Wise, 1988). The Barnes delta is thought to have developed while in contact with ice filling the basin to the north. The lack of any other outwash deposits suggests that the entire basin extending north from the Holyoke Range to Greenfield was freed of ice in a single stagnation event. A second stagnation event would then account for the removal of ice from the Montague basin. With the removal of the ice from the Montague basin, the ice edge in the Connecticut Valley lay near the present northern edge of the Millers River delta, and so the delta began to develop. This edge corresponds to the position of a buried bedrock obstacle (Tighe and Bond, 1988) which we infer impeded the movement of the glacial ice down the Connecticut Valley.

The melting of the ice to the north of the Millers River delta caused the cessation of major sedimentation on the delta (there may have been some minor continued sediment input from the Millers River Valley and smaller feeder streams) and created northern Lake Hitchcock. Although generally considered a continuation of the main body of Lake Hitchcock

(Stewart, 1961; Koteff and others, 1988), northern Lake Hitchcock was separated from the main body of the lake by the barrier of the delta. It is probable that a short river flowed from northern Lake Hitchcock to the main body of the lake.

## GEOLOGY OF DELTA

The Millers River delta consists of three separate morphologic portions: Montague Plain, Turners Falls, and French King terrace. The Millers River separates Montague Plain from French King Terrace, and the bedrock barrier of the Mineral Hills and Willis Hill separate the Turners Falls portion from Montague Plain. Seismic surveys (Weston Geophysical, 1966; Tighe & Bond, 1988) indicate the thickness of the deltaic sediments ranges up to about 300 feet within the pre-glacial Connecticut River channel underlying French King Terrace and Montague Plain. The present channel of the Connecticut River crosses the bedrock barrier and flows over Turners Falls, probably reflecting historical flow patterns across the top of the delta.

The leading edge of the delta is morphologically distinct along the edges of Montague Plain (Montague-Turners Falls Road follows the base of the delta front), but is relatively indistinct along the Turners Falls portion. Jahns (1966) attributed the gentle slopes along the edge of the Turners Falls portion to beach erosion; we, however, prefer to attribute it to delta front collapse caused by groundwater percolation through the delta from the Connecticut River following the lake level drop to the south and before the incision of the delta by the river. The evidence, however, is insufficient at this time to distinguish between these two hypotheses, although the lack of beach erosion along the edges of Montague Plain would tend to detract from Jahns' (1966) explanation.

Kettles are prominent features in both Montague Plain and French King terrace. In Montague Plain there is a line of kettles extending from the Millers River gorge to the edge of the delta. The linear depression created by these kettles has been called a former channel of the Connecticut River (Emerson, 1898), but Jahns' (1966) mapping of late-stage channels across Montague Plain shows that the last flow directions emanate from the mouth of the Millers River Valley and cross this line. We attribute this line of kettles to ice caught in a lateral/medial moraine buried beneath the delta sands. The multiple large kettles in the French King terrace are consistent with a nearby ice margin.

The Connecticut River valley in Massachusetts north of the French King terrace contains only small kame deltas and river terraces that grade to the Lily Pond channels (Campbell and Hartshorn, 1980). There is no evidence of any stream deposits grading to the north end of the delta. The sediments at the north end of the delta contain both thrust and collapse

features. Since flow indicators in the topset beds of the French King terrace show southward stream flow, we propose that this terrace was bounded to the north by ice.

## POST-DEPOSITIONAL HISTORY

Following the deposition of the Millers River delta, the ice margin continued to retreat northward up the Connecticut Valley. The delta filled the junction between the narrow valley to the north and the broad valley to the south, thereby dividing Lake Hitchcock into two lakes. Because the outlet to the lake system lay to the south in central Connecticut, water should have flowed across the top of the delta deposits from the northern lake into the southern lake.

When the morainal dam in central Connecticut was breached, the level of southern Lake Hitchcock dropped relatively rapidly. The lake level to the north of the Millers River Delta, however, remained high until the river breached the delta deposits. This breaching can be expected to have occurred by upstream nickpoint migration. Changes of the channel direction may have occurred by groundwater flow through the deltaic sediments and upgradient migration of spring-fed streams until river capture occurred. We hypothesize that this may have occurred in the vicinity of Turners Falls. The original channel through the White Ash Swamp channel in Greenfield (Jahns, 1966) was captured and diverted to the present channel by groundwater seeping to the old delta front.

As the river incised into the delta, it finally encountered a bedrock barrier commonly called the Lily Pond barrier. The highest notch through this barrier was initially encountered, but the river then jumped to a lower, upstream notch and then finally to its present, downstream notch. These various river levels are clearly recorded to the north by river terraces grading to each of these notches (Jefferson, 1898; Campbell and Hartshorn, 1980). The causes of the changes to new channels are again attributed to groundwater seepage through the lower notches leading to river capture.

The timing of capture of the Millers River is not clear. Either pre-lowering or post-lowering scenarios are possible. The Fall River was probably diverted to its present course to Turners Falls when the Connecticut River was diverted away from White Ash Swamp channel.

## CONCLUSIONS

The Millers River Delta was apparently formed in contact with stagnant glacial ice in the Connecticut River Valley. Thus, it is a classic lacustrine ice-contact morphosequence.

Following ice retreat northward in the Connecticut Valley, this delta divided lake Hitchcock into two distinct water bodies. As the lake level dropped in the southern

portion, the delta delayed the drop to the north.

The incision of the river into the delta proceeded in stages. The procession of these stages was influenced by bedrock topography and groundwater flow patterns.

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Bernardston
Greenfield
Gill
Erving
Montague
Turners Falls
Millers Falls
Montague City
Lake Pleasant
Riverside
Start
Lunch

White Ash Swamp
Start
8
Barton Cove
Barrier
CONNECTICUT
RIVER
Delta Top
Collapsed
Rock
3
Rock
Delta Top
5
Millers Falls
4
1
Delta Top
2
Kettles
MONTAGUE
Foreset
MONTAGUE

# ROAD LOG

Meeting point: Turners Falls Overlook on Route 2 just west of Turners Falls Bridge.

This vantage point shows the present gorge of the Connecticut River where it has cut totally through the deltaic sediments down to the underlying bedrock. The bedrock falls serve as baselevel for the upstream portion of the Connecticut River. The tops of the surrounding hills are at or near the level of the delta top.

0.0 Proceed East on Route 2.

0.1 Turn Right across Turners Falls Bridge.

0.7 Turn Left at light onto 3rd Street.

The top of this hill is the delta top. This hill slope was created by the incision of the Connecticut River into the delta (Jahns, 1966).

1.5 Bear Right onto Montague Road.

2.3 Proceed Straight.

The gentle slope marks the front of the delta in this area. We hypothesize that this gentleness is due to collapse under artesian groundwater conditions following the lake level drop to the south; Jahns (1966) considered this to be a beach.

3.3 Note the steep delta front to the left. We are riding along the base of the delta front. When the vegetation is absent, channels can be identified by their coarser grain-size.

4.0 Turn Left.

4.1 STOP 1

Within the gravel pit to the left we will see topset and foreset beds of the delta. These deposits indicate flow toward the delta front and do not indicate beach reworking.

Proceed onto the delta and continue straight. Note the gentle topography off to the sides created by the late channels across the delta; the road is very nearly parallel to flow direction.

4.8 Continue Straight.

5.9 Turn Right onto paved road.

6.9 Turn Left onto Green Pond Road.

7.2 STOP 2
The depressions to the north and south are kettles. They are part of a line of kettles that we believe mark a marginal morraine buried beneath Montague Plain. Lake Pleasant to the south occupies the largest kettle in western Massachusetts (Larsen in Tighe & Bond, 1988).

Proceed.

7.5 Turn Left onto dirt road.

8.0 Turn Right onto paved road.

8.9 Turn Left.

9.4 STOP 3
Much of the top of the delta is covered with dune deposits. This is especially true around the Turners Falls Airport which we are next to. Here we will look at some stabilized dunes. Just to the north lies the present Connecticut River.

Turn Around and proceed eastward to Millers Falls.

11.1 Turn Left and cross the Millers River. The River has cut across the delta to join the Connecticut River about a mile downstream.

11.5 Turn Right toward Route 2.

12.1 Proceed East on Route 2.

12.7 STOP 4
The landfill to the right is used for sewage sludge from the Erving Paper Mill. At the east end of this pit there are some excellent examples of till overlain by lake clays. These clays must have been deposited in an ice-marginal lake. No obvious river terraces grading to the delta have been mapped in the Millers River Valley.

Turn around and proceed back west on Route 2.

13.6 Turn Right toward Route 63 north.

13.9 Turn Right onto Route 63.

15.1 As we pass under the power lines, we reach the north end of the delta. Note the lack of stream terraces grading to the north end of the delta.

16.0 Turn Right into Northfield Mountain reception center.

16.3 LUNCH

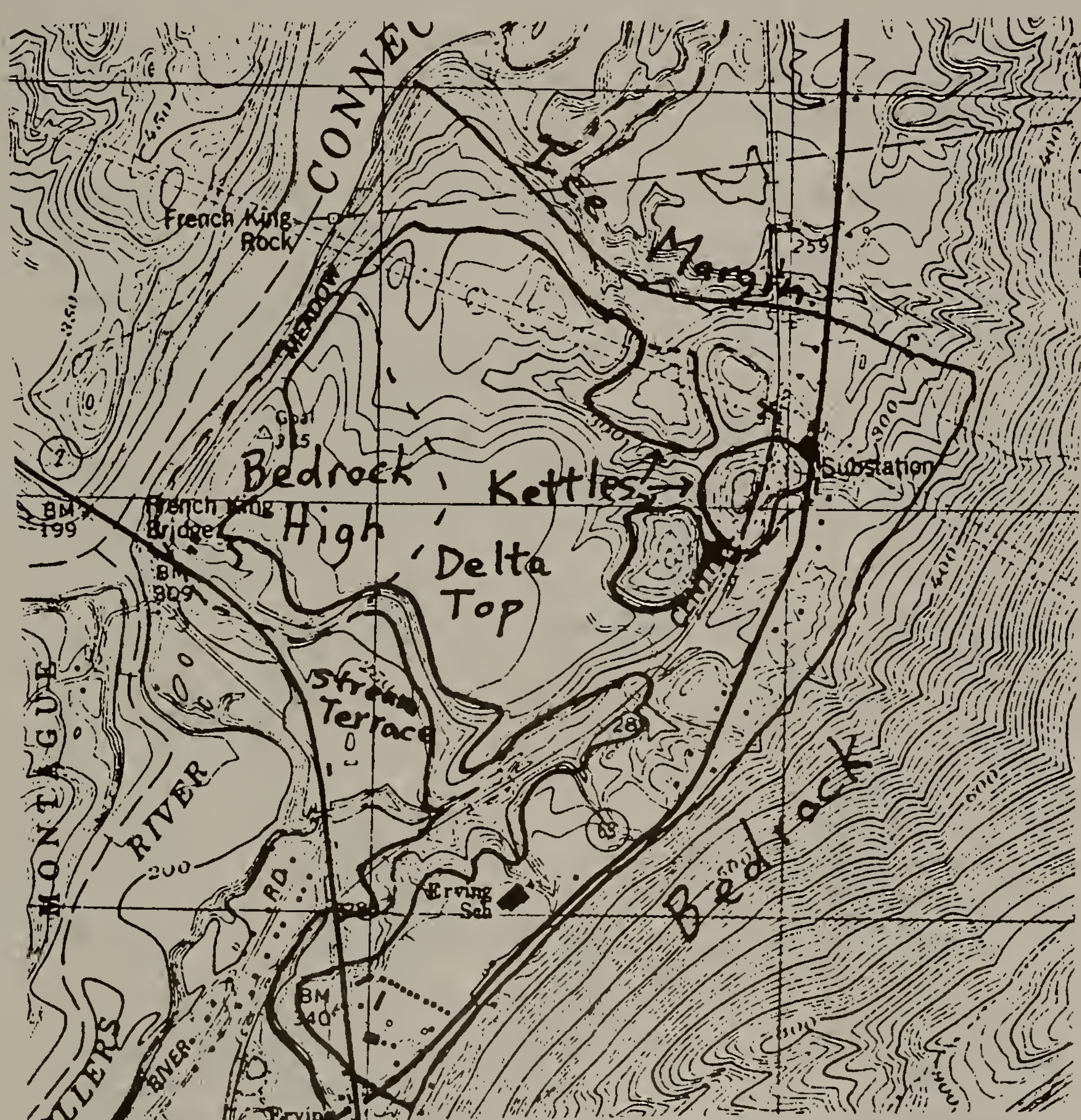
After lunch return to Route 63 and proceed back south toward delta.

7.4 STOP 5

Proceed back along road about 50 yards and follow trail back into woods to train tracks. Cross tracks and proceed to trail junction. In front of us lies a large (about 10 acre, >50 ft deep) kettle. This kettle is essentially dry. Just behind us lies another large kettle transected by the tracks. A third large kettle lies a short distance to the northwest. This large kettle is scheduled to be filled during construction of the Erving Industrial Park (Tighe & Bond, 1988).

Proceed back to the tracks and follow them northward across the causeway to the cut. Excavations along this cut have found both thrust and collapse structures. Sedimentary structures indicate flow toward the south.

Return to cars and proceed south.



18.5 Turn Right.

18.5 Turn Right onto Route 2 west.

19.5 Connecticut River.

19.6 STOP 6
Walk back to French King Bridge. This provides a good vantage for viewing the geomorphic relations around the north end of the delta. French King Rock in the Connecticut River just to north of the bridge is a large boulder.

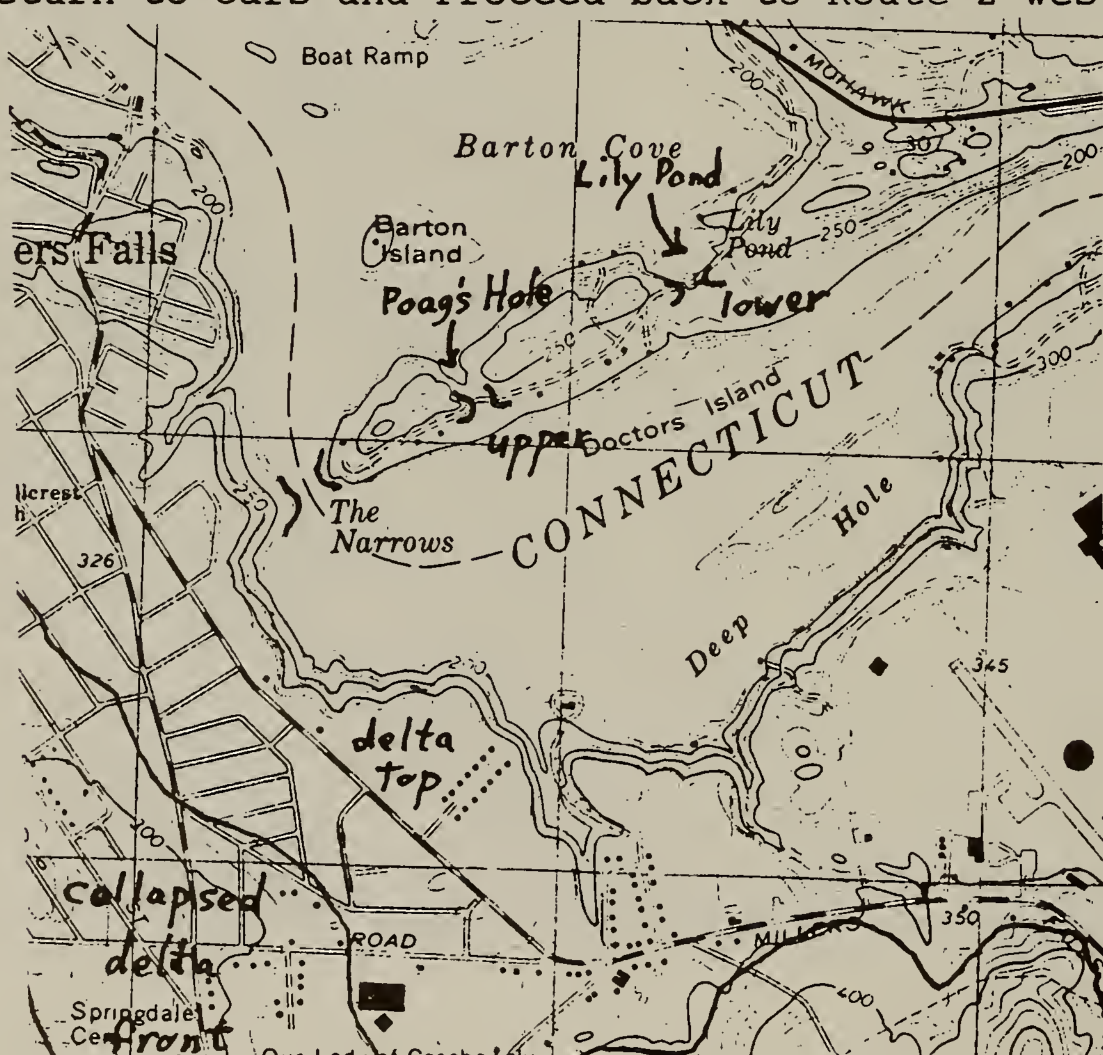
Return to the cars and proceed west.

21.9 Turn Left.

22.3 STOP 7
This is the Lily Pond Barrier. As we walk down to the notches, stream terraces can be seen along the east side of the peninsula. The first notch is the lower Lily pond notch. The plunge pool was the site of Lily Pond (Jefferson, 1898) before the water level was raised at the Turners Falls dam. The second notch is the the upper Lily Pond notch, where the plunge pool was called Poag's Hole. The end of the peninsula borders the present notch called the Race.

Directly beneath the lip of the upper Lily Pond notch is a Mesozoic stratum comprised of rip-up clasts and mudballs. This is thought to record Mesozoic storm sedimentation.

Return to cars and Proceed back to Route 2 west.



23.5 Proceed straight.

23.7 To left is starting point for trip.

24.3 STOP 8

The sand pit to the left contains excellent exposures of foreset and topset beds indicating southwestward current directions. The boundary between topset and foreset can be seen, and it should be noted that the boundary is actually a zone indicating that sedimant is delivered to the delta front as packets which then slump down the slope. The existence of this boundary zone makes determination of water levels from the boundary highly questionable, at least with the degree of accuracy Koteff and Larsen (in Koteff and others, 1988) have claimed.

Return to cars. Route 2 west toward I-91 follows the White Ash Swamp channel until it reaches Route 5.

END OF TRIP

# THE QUATERNARY GEOLOGY OF THE UPPER ASHUELOT RIVER, LOWER COLD RIVER, AND WARREN BROOK VALLEYS OF SOUTHWESTERN NEW HAMPSHIRE

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## INTRODUCTION

The Ashuelot River, Cold River and Warren Brook drainages are a part of the westward draining tributary systems of the Connecticut River in southwestern New Hampshire (Figure 1). The Ashuelot River flows to the south-southwest from its headwaters west of Lovewell Mountain and through Surry Mountain Lake and the town of Keene before heading west and entering the Connecticut River at Hinsdale. Warren Brook drains west from the area of Lake Warren to Alstead where it enters the Cold River. The Cold River flows west from its confluence with Warren Brook and enters the Connecticut River north of Walpole. Bedrock in the region is primarily pegmatite-bearing schist, gneiss, and granite (Robinson and others, 1979). The lineation, foliation and gross structure of the rocks, as well as differences in weathering and resistance to glacial erosion, control the shape of the topography in the area.

All three drainage systems have been influenced by glaciation, and ice-dammed lakes that provided local base level controls. Weathering of bedrock prior to at least one glacial advance in the region is also recorded in the Cold River and Warren Brook valleys. The Quaternary features and events that occurred in the three valleys are discussed below in order of their relative ages from the oldest to the youngest. However, it is not clear how all of the glacial units in the region fit into a regional glacial chronology.

## WEATHERED BEDROCK

Several exposures in the Warren Brook and Cold River valleys reveal saprolite and weathered rock that are overlain by either (glacio-?) fluvial gravel or till or are draped by sub-till glaciolacustrine sediments (Figure 2). Saprolites are generally found in schist, while granitic gneiss may be partially weathered beneath overlying glacial sediment. Weathering in saprolites includes the alteration of feldspar to kaolinite and an orange-red color produced by the formation of iron and manganese oxides. In the Warren Brook region, depths of weathering are usually no more than 2 meters but are known to exceed 7 meters in schist. Evidence that saprolite weathering is not simply the result of post-glacial groundwater alteration is that deformed and smeared, weathered schist occurs in the base of till in the Warren Brook area. In addition, unoxidized glaciolacustrine strata have been found draped over colluviated schist that is overlying partially

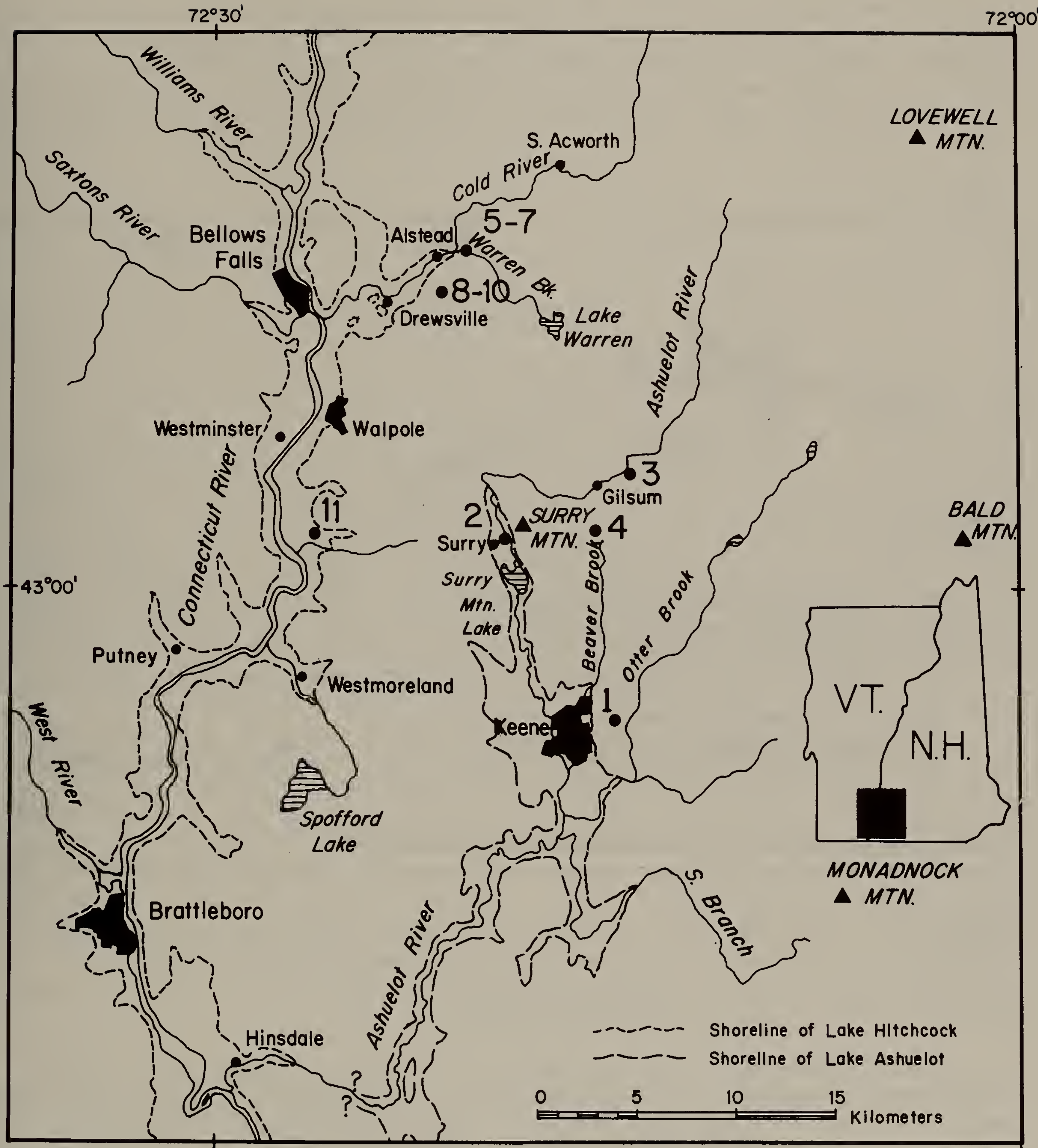


**Figure 1** - The Connecticut valley region of southwestern New Hampshire showing the Cold River, Warren Brook and the Ashuelot River. The shorelines of Lake Hitchcock and Lake Ashuelot are also shown. Field trip stops are numbered.

weathered schist that has not been colluviated. In both of these cases bedrock was weathered prior to at least one glaciation and lake impoundment. At Cock Hat Hill (Figure 2, Stop 7), erosion by ice-marginal drainage appears to have taken advantage of non-resistant saprolite in the cutting of meltwater channels. The upward increase in weathering intensity, from unweathered to highly altered bedrock, seen at some weathered rock localities is consistent with the interpretation of these exposures as buried soils.

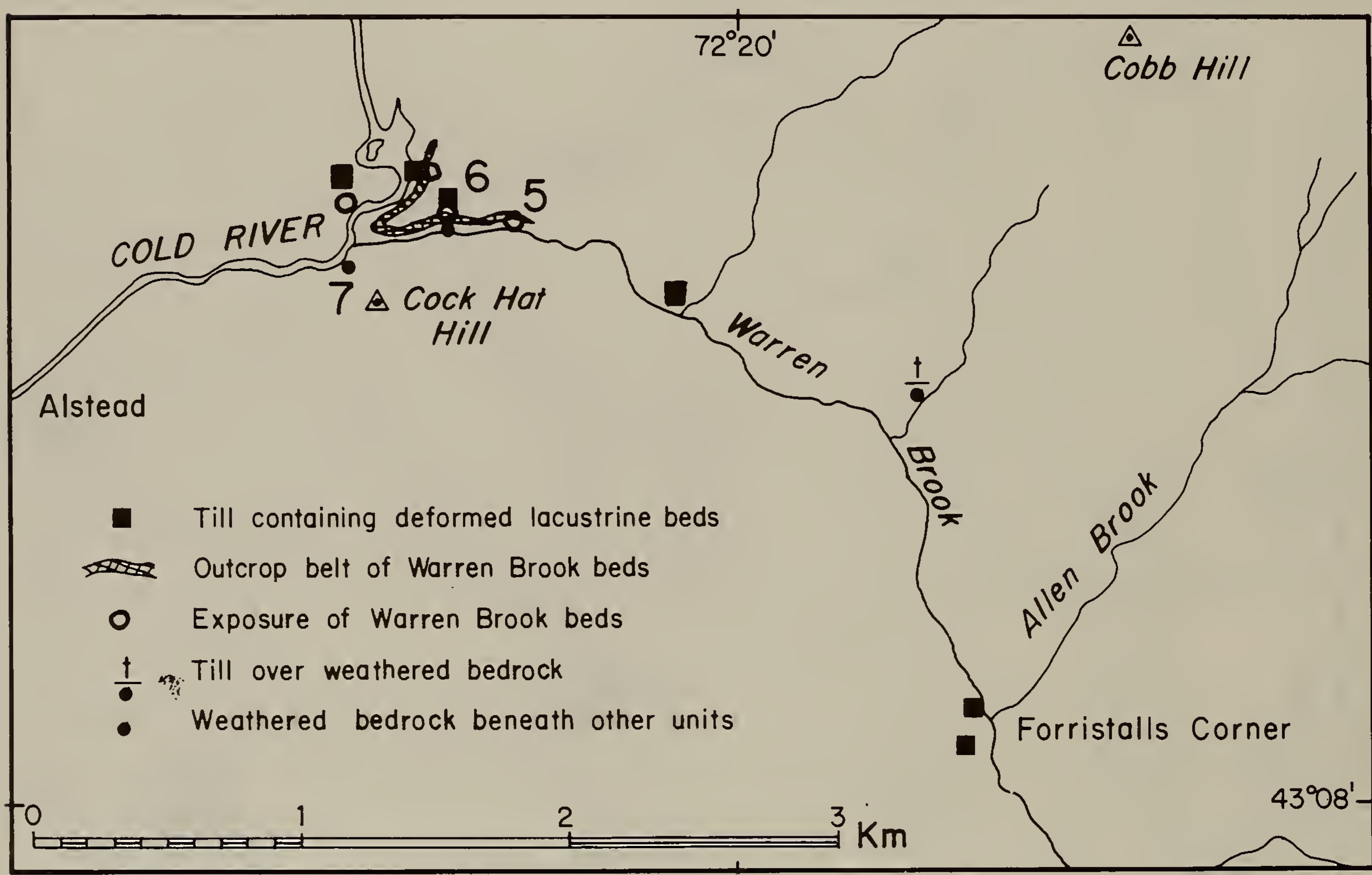


**Figure 2** - Exposures of the Warren Brook beds and weathered bedrock in the Warren Brook valley between Alstead and Forristalls Corner. Numbers indicate field trip stops. Base is Bellows Falls (1:25,000) Quadrangle.

Other exposures of saprolite or "rotten rock" have been reported elsewhere in New Hampshire (Goldthwait and Kruger, 1938) and in other New England states and Quebec (LaSalle and others, 1985; J.P. Schafer, pers. com.). Because of their intensity and depth of weathering it seems unlikely that saprolites could be attributed to the last episode of interglacial weathering and they may be as old as Tertiary (LaSalle and others, 1985). However, it is difficult to make the same statement about saprolite and weathered rock in the Warren Brook valley because weathered rock thicknesses have not been found greater than 2 meters except in easily weathered schist.

## SUB-TILL LACUSTRINE SEDIMENT (Warren Brook beds)

At five exposures in the Warren Brook and Cold River valleys near Alstead, lacustrine sediment has been found beneath till (Figure 2). For the sake of communication the lacustrine units which occur beneath till in the Warren Brook area will be referred to as the 'Warren Brook beds'. The Warren Brook beds are entirely clastic and mostly light gray, laminated, fine sand and silt with dropstones and occasional rippled, micaceous, and sometimes gravelly, coarse to medium sand beds. Some sandy beds and partings are composed entirely of brownish-orange mica and schist fragments which appear to be derived from local weathered bedrock. None of the units contain clay beds which probably indicates rapid and nearly continuous deposition into the lake. The sinusoidal trace fossils of nematode worms, common in glaciolacustrine sediments elsewhere in New England, have been found in laminated silt and fine sand beds exposed along Warren Brook (Stop 6). The upper parts of exposures of the Warren Brook beds display evidence of glaciotectonic deformation in the form of fractures, small thrust faults, and smearing and homogenization of beds which are truncated beneath overlying till. Till immediately above the Warren Brook beds is sparsely stony, very silty and contains deformed lacustrine sediment and sheared weathered bedrock. The Warren Brook beds are silty, and where they have not been loosened by groundwater seepage or exfoliation, they are extremely compact as a result of overriding ice. The Warren Brook beds are known to occur as far east as the area of South Acworth in the Cold River valley.

For lack of an alternative mechanism of damming the Warren Brook and Cold River valleys, the Warren Brook beds appear to have been deposited in a short-lived proglacial lake impounded by the advance of lobate ice in the Connecticut Valley. Supporting evidence from the Warren Brook beds for proglacial deposition in cold water are the exclusively clastic character of the sediment, dropstones, the traces of nematode worms, and abundant silt (rock flour).

## ICE FLOW DIRECTIONS

Numerous outcrops of quartz-bearing pegmatite and vein quartz in schist bear grooves, striations, and crag and tails from which late Wisconsinan ice flow can be determined. Striations in the Connecticut Valley and adjacent uplands reflect ice flow controlled by local topography and regional ice flow to the southeast. Regional ice flow appears to be about South 5 to 15 degrees East and is best seen on the broad low-relief uplands in the eastern Walpole Quadrangle (East side of Figure 3). This pattern is repeated to the north in the Bellows Falls Quadrangle. Ice flow in the Connecticut valley is predominantly parallel to the trend of the valley axis and reflects the confinement of ice by local topography. In Vermont (southwest corner of Figure 3) ice flow is parallel to the trend of bedrock hills and is South 10 to 25 degrees West. Variations in striation directions appear to be the result of deflection of ice flow by local topography.

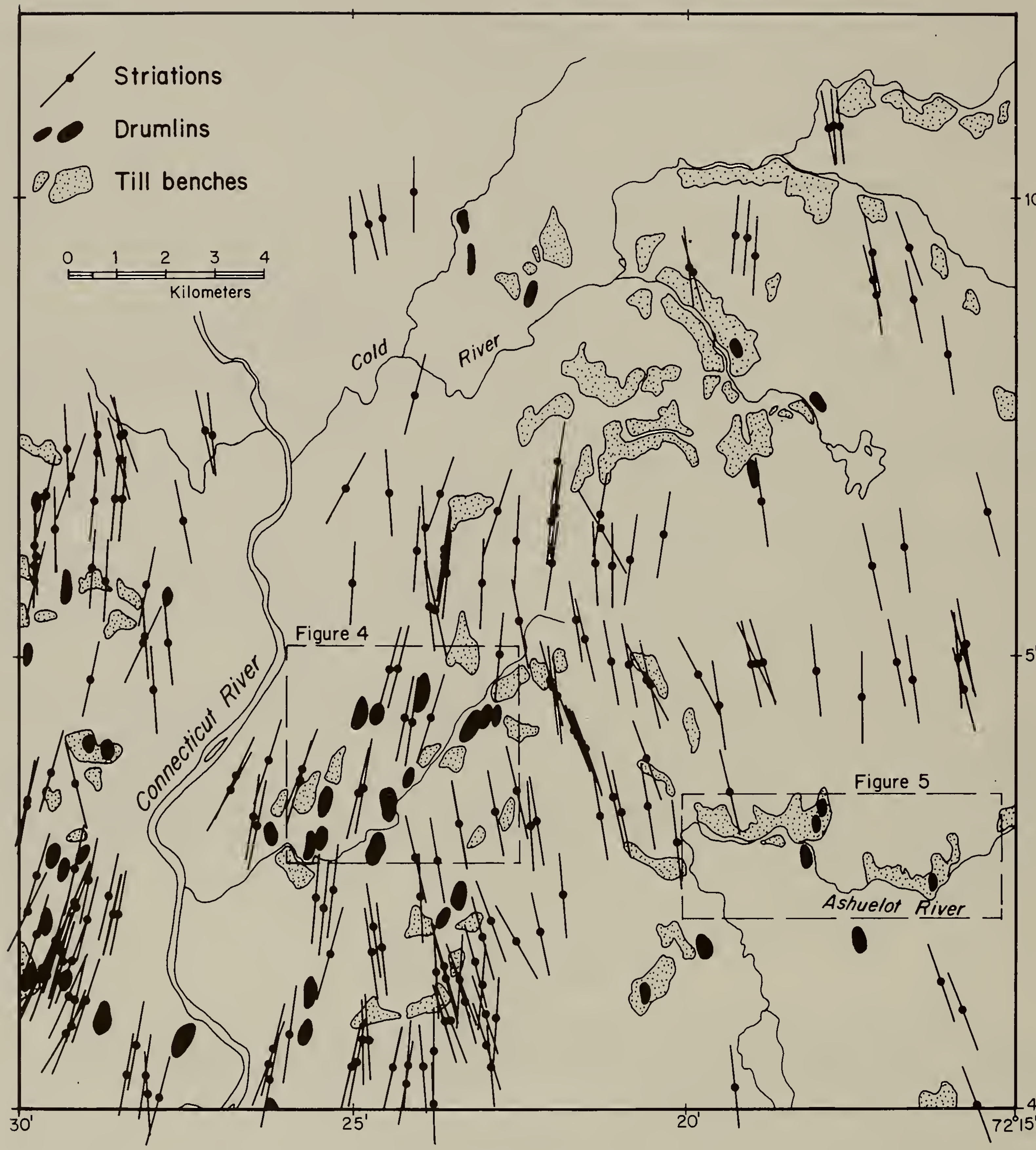


**Figure 3** - Occurrence of drumlins, till benches and ice flow indicators in the Walpole and southern Bellows Falls (1:25,000) Quadrangles.

Drumlin axes are parallel to striations on adjacent bedrock hills and it appears that drumlin forms are the product of late Wisconsinan ice flow. Existing striation data does not confirm or eliminate lobate flow of ice in the Connecticut Valley. No multiple striation localities have been found in the Walpole Quadrangle that suggest changes in ice flow direction during deglaciation.

## GLACIAL DIAMICTON FACIES

Several glacial diamicton facies have been found in the Connecticut valley and adjacent uplands of southwestern New Hampshire (Table 1). (Note: 'Diamicton' is here used to indicate any poorly sorted, matrix-supported conglomerate, regardless of origin.) Diamicton lithologies range from sparsely stony, clayey and silty diamictons to very stony, sandy diamictons. Diamictons may be oxidized, stained, or non-oxidized, and may be massive, layered, or contain sorted sediment in the form of beds or lenses. All the diamictons contain striated and glacially shaped clasts indicating that they are in some way related to glacial sedimentation. Diamicton facies have been catagorized as end members (Table 1) but variations occur within these catagories. Gradational combinations of end-member facies have not been recognized but continued field work will be needed to affirm that facies are not gradational.

The wide variety of diamicton facies may be explained in several ways. Different diamicton facies may be the result of:

1) deposition in distinct subglacial, supraglacial, and proglacial environments;

2) deposition during changes in the character of subglacial environments, particularly with respect to the amount of water at the base of the glacier in places where rates of subglacial melting, release of englacial debris, and ice flow vary over time;

3) depositional responses to differences in provenance which are a function of the types of bedrock and surficial deposits that a glacier overrides; and

4) deposition during different glaciations when the conditions listed in the items above may have varied.

The mechanisms above may have operated simultaneously which complicates an identification of genetic types of diamictons based on lithology. Without consistent field criteria for the recognition of diamicton facies as either: 1) proglacial sediment flow or pelagic rainout deposits, 2) supraglacial meltout till, 3) subglacial meltout till, or 4) lodgement till; the task of assigning diamictons to ages or glaciations is difficult.

A review of the lithology and occurrence of multiple till units in other parts of New England and New Hampshire, and the possible correlation of these units with specific glaciations is given by Koteff and Pessl (1985). A clayey and silty lower till that is early Wisconsinan or older and a sandy upper till which is late Wisconsinan in age have been recognized elsewhere.

Table 1. -- Summary of diamicton lithofacies recognized in the Connecticut River drainage basin of southwestern New Hampshire and adjacent Vermont.

| END-MEMBER DIAMICTON FACIES | MAXIMUM OBSERVED THICKNESS | POSSIBLE GENETIC INTERPRETATIONS |
|---|---|---|
| 1. Stony, sandy diamicton interbedded with sorted sand and gravel that is crudely to well stratified. | 3-4 meters with interbedded sand and gravel. | Proglacial sediment flow or slumped supraglacial deposits (flowtill) and stratified deposits. |
| 2. Mostly massive to very crudely layered, stony, sandy diamicton with occasional to numerous subhorizontal partings of silt to medium sand. Discontinuous layers or pods of muddy diamicton and lenses and infilled fractures of sand may occur. | Up to 3 meters (limit of exposure) | Proglacial sediment flow deposit, subglacial meltout till, or lodgement till. (Are the sandy partings a shear structure ?) |
| 3. Stony to sparsely stony, muddy diamicton with crude layering interbedded with laminated beds of clayey silt | 3 meters, top interbeds with lacustrine sediment, bottom rests on facies 4. | Subaqueous proglacial sediment flow deposits (flowtill) or pelagic rainout deposits interbedded with lacustrine sediment. |
| 4. Dark greenish gray (5GY 4/1), non-oxidized, compact, sparsely stony to stony, muddy diamicton. | Up to 30 meters (limit of exposure) | Lodgement till |
| 5. Olive or yellowish gray (5Y 4/3-4), oxidized, compact, sparsely stony to stony, muddy diamicton. | Up to 10 meters (limit of exposure) | Lodgement till (Higher chroma than facies #6 appears to be related to recent soil oxidation.) |
| 6. Olive gray (5Y 4/2), oxidized or stained, compact, stony, muddy diamicton. Blocky structure, black (10YR 2/1), iron-manganese stains on sand grains, pebbles and fracture planes. | Up to 5 meters (limit of exposure) | Lodgement till (Diamicton that most resembles the oxidized lower till elsewhere in New England. Is discoloration related to pre-late Wisconsinan oxidation ?) |

Lithologic correlation of some of the southwestern New Hampshire diamicton facies with either the lower or upper tills elsewhere in New England is tempting. However, in light of the preliminary nature of investigation in southwestern New Hampshire, the degree of variability of late Wisconsinan diamicton (due to provenance changes), and the pausity of clear structural or stratigraphic relationships between diamictons in single outcrops, correlations at this point seem tenuous.

## DIAMICTON PROBLEMS

Several problems, that may not be unique to southwestern New Hampshire, must be dealt with before a correlation with other areas can be formulated. These problems were not only realized during field mapping over the last three years but are also derived from discussions with others who have more experience with the till stratigraphy of New England.

Problem 1: Using Lithology as a Correlation Tool
Within southwestern New Hampshire and adjacent Vermont, late Wisconsinan till does not have a consistent litho-type. Late Wisconsinan till may be either stony and sandy, as is the case on the uplands near Keene and in Hammond Hollow near Gilsum, or muddy with fewer stones. Muddy, sparsely stony, late Wisconsinan till occurs in the Warren Brook and Cold River valleys, where till overlies the silty Warren Brook beds, and throughout the region in places down flow from schist and phyllite. It is important to note that regional ice flow to the southeast would put areas of soft phyllite, slate and schist in Vermont up ice from the Walpole area. Sheared weathered rock is also found in till of the Warren Brook valley, and near Westminster, Vt. a very sparsely stony, clayey till appears to correspond to very friable underlying schist and phyllite. While a muddy matrix is a general characteristic of the lower till in other parts of New England, grain size may not be used to consistently identify till in southwestern New Hampshire and adjacent Vermont that is pre-late Wisconsinan in age.

Problem 2: The Origin of Discoloration in Diamictons
Discoloration is here used to describe any color which was not inherited by a diamicton at the time of deposition. Discoloration can be the product of oxidation of a diamicton or the staining of a diamicton by oxide or hydroxide precipitation. In some situations it is possible to identify the cause of discoloration as Holocene soil formation that has produced a surface oxidized zone. However, in some outcrops discoloration has been seen to depths of at least 10 meters, may occur beneath non-oxidized deposits, and is not clearly related to recent soil formation. Manganese stains on sand grains and on fractures and a high degree of consolidation may be related to cementation by iron-manganese oxides and hydroxides. It is difficult to determine whether this form of discoloration is entirely related to surface weathering at some time in the past, or is partly the result of groundwater oxidation or the downward recession of the water table. Disseminated fine-grained pyrite, in diamictons

derived from pyrite-bearing bedrock, is easily oxidized and probably plays a role in allowing subsoil oxidation of diamictons and lacustrine beds.

Problem 3: The Identification of Small-Scale Shear Structures

Structural features at the contact between till units has allowed the distinction between the upper and lower tills elsewhere in New Hampshire (Kotef and Pessl, 1985). Superposition of unoxidized over oxidized thrust blocks of the lower till, pieces of the lower till in the upper till, and a shear zone separating the two till units are characteristic of the upper till/lower till contact. Large-scale structural features, like those above, provide strong evidence of deformation of the lower till during the deposition of the upper till. However, the interpretation of small-scale features, especially in sandy diamictons, as shear structures is not straight forward. Small-scale features are important to the identification of lodgement till versus subglacial meltout till. Several types of sandy and silty partings, fracture fillings and lenses may or may not be related to shearing. Alternative explanations for some of these features may be: 1) crudely layered palimpsest structures related to englacial debris layering that persists after meltout, 2) intratill partings and lenses deposited in channels or water films at the sole of a glacier, 3) sandy partings and fracture fillings resulting from the infilling of dewatering veins, and 4) horizontal fissility related to exfoliation or unloading. The identification of small-scale shear structures should be limited to features with clear evidence of deformation or displacement.

An additional structural feature, that appears to be a characteristic of muddy till, is a platy or blocky structure that is frequently stained with iron-manganese oxides. The origin of this structure is not known and it does not appear to be fissility related to shearing which would be more horizontal. It is not known whether the blocky structure that is common in the lower till elsewhere in New England is related to oxidation or staining, or whether it is characteristic of the deformation of an older till by subsequent glaciation. It is also not known whether the blocky structure is restricted to the lower till or can occur in muddy diamictons of late Wisconsinan age.

## DRUMLINS AND TILL BENCHES

In the Connecticut valley and adjacent hilly regions of Vermont and New Hampshire, two types of landforms, drumlins and till benches (Figures 3, 4 and 5), are composed of thick (up to 30 meters exposed) till. The till in both features is always compact and muddy (facies 4, 5 and 6 of Table 1) and may be unoxidized (facies 4) or oxidized (facies 5 and 6) in thick exposures. Till in drumlins and lee-side till benches contain numerous boulder and pebble lines, horizontal partings from which soil moisture seeps, and pockets, lenses and beds of stratified sediment (Stop 8). The clast lines and stratified sediments (intratill beds) are believed to be the result of deposition in

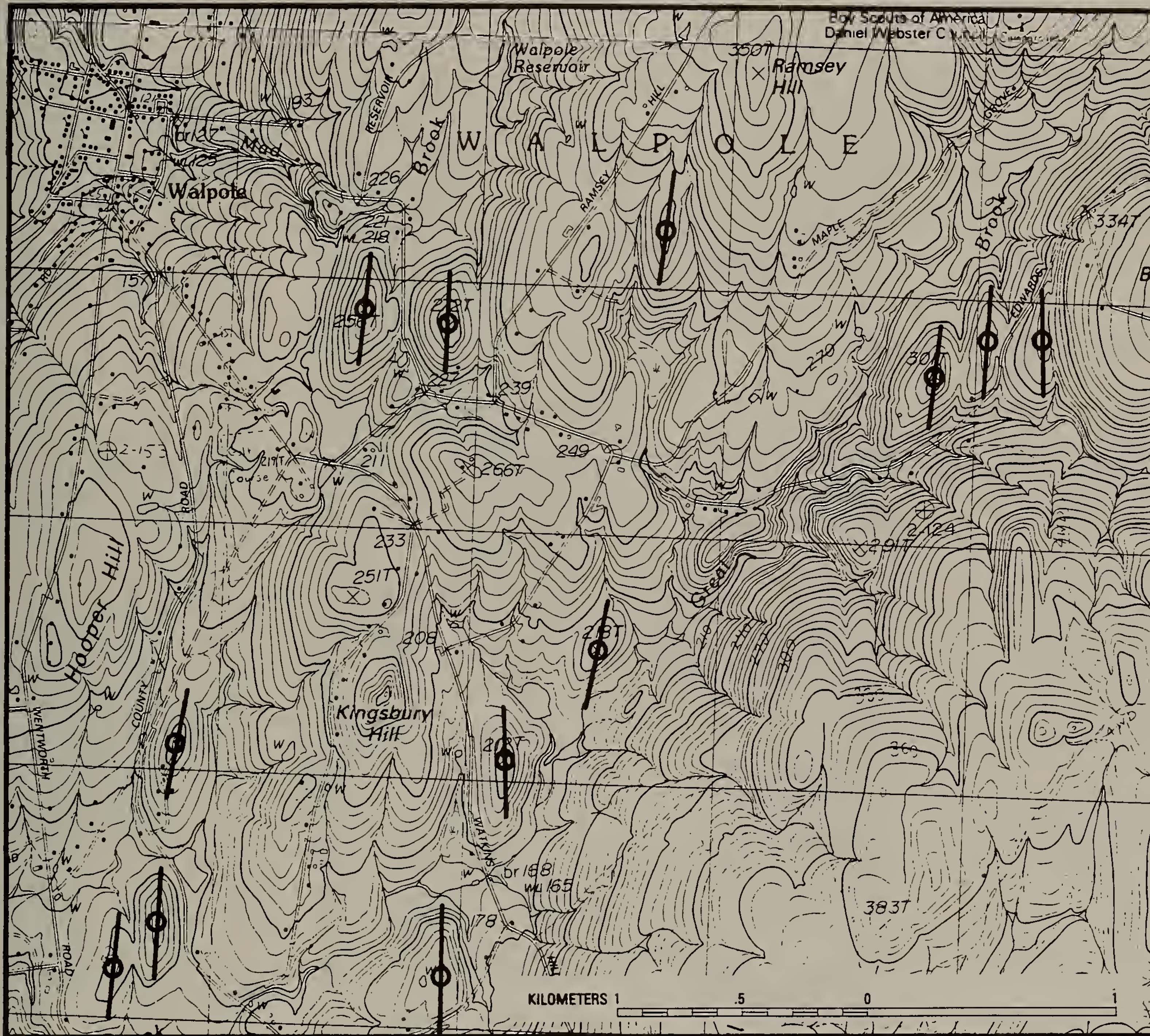


**Figure 4** - Examples of drumlins in the vicinity of Walpole, N.H. Base is Walpole (1:25,000) Quadrangle. Contour interval: 6 meters. See Figure 3 for location.

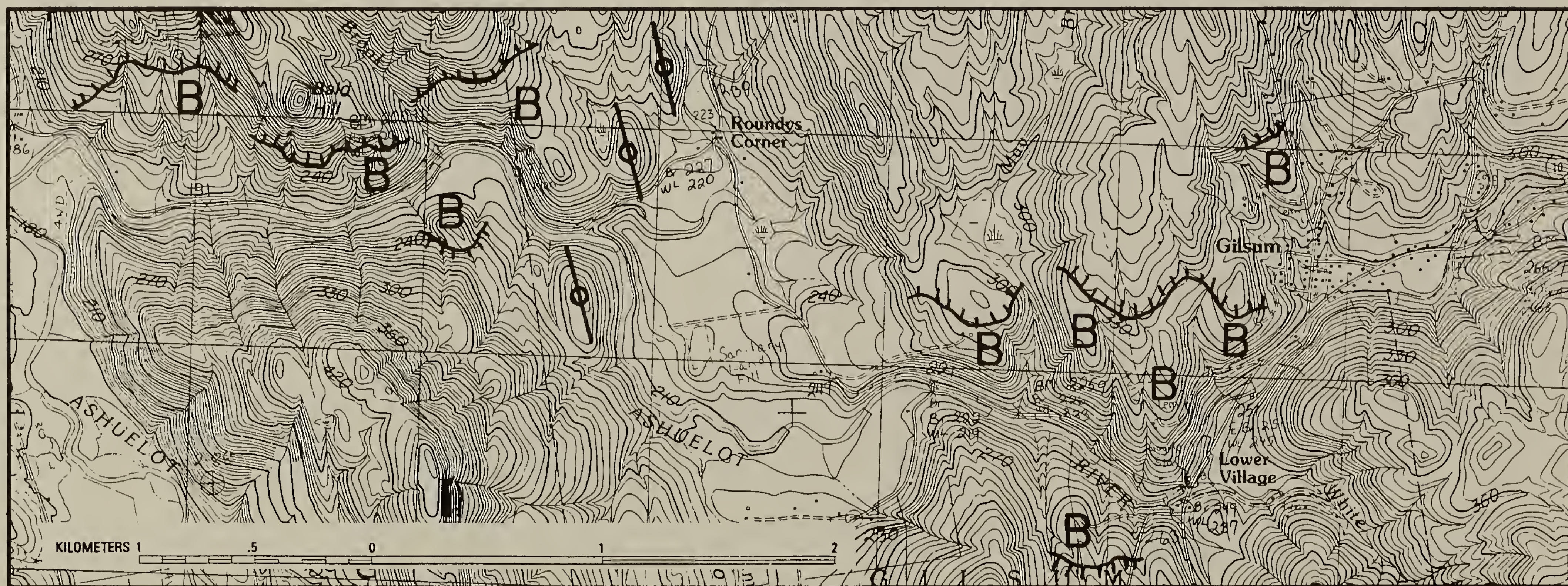


**Figure 5** - Till benches and drumlins in the Ashuelot River valley west of Gilsum. Till benches are marked with B's. Dark lines with tic-marks indicate the position of a break in slope that marks the upslope extent of benches where they are adjacent to thin drift or bedrock. Base is Walpole (1:25,000) Quadrangle. Contour interval: 6 meters. See Figure 3 for location.

water-filled, subglacial openings formed by cavitation on the lee sides of bedrock highs. Cavitation deposits are generally not laterally extensive, vary over a wide range of grain sizes, and include mass movement, rainout, and current-derived beds. Subglacial cavitation deposits include rhythmic, laminated silt, clay and sand beds that can be confused with overridden proglacial lacustrine sediment. Subglacial cavitation deposits invariably contain abundant non-lithified diamicton drop sediment ('drop clots') derived from the meltout of englacial debris at the base of the glacier. Proglacial lacustrine sediments can be recognized by the occurrence of sinusoidal trace fossils of nematode worms.

Drumlins are most common in north-south trending basins and appear to be in the lee of irregular bedrock highs or other drumlins (Figures 4). It is not known whether the drumlins are depositional or erosional landforms. Till in drumlins is exposed only where stream erosion since the last glaciation has cut into these features and it is not known whether drumlin till in the region is late Wisconsinan.

Till benches (Figures 3 and 5) occur on the north and south sides of bedrock uplands in east-west trending valleys. Their occurrence indicates that till benches are probably the result of stoss- and lee-side till deposition or protection from erosion during the last glaciation. In some places a subtle streamlining of bench surfaces, that is parallel to drumlin orientations and striations, suggests that they may be related to drumlin formation. The gentle bench slopes, which abruptly end at steep valley faces, suggest a once continous infilling of the valley in which they occur and that they were probably incised by streams since the recession of the last glacier. Exposures of till in the benches occur along main valleys where they are being undercut by streams and in gullies incised by small streams that cross from bedrock uplands on one side to valley floors on the other side of the bench.

## ICE RECESSION IN THE UPPER ASHUELOT VALLEY

As a consequence of the westward course of the Ashuelot valley from east of Gilsum to near Surry, and earlier ice recession to the east than to the west, a series of lakes were dammed by ice west of Gilsum. An ice-dammed lake drained across a col on the upland south of Gilsum at Bingham Hill (345 meters, Stop 4) into the Beaver Brook valley (Figure 6A). A col on Surry Mountain (320 meters) west of Lily Pond (at least 25 meters deep) later served as a spillway and plunge pool that allowed drainage into the lower Ashuelot Valley (Figure 6B). Ice recession eventually let water escape around the north end of Surry Mountain at an elevation of 200-260 meters and deltas were deposited in the Ashuelot valley north of Hammond Hollow (Figure 6C). Evidence of glacial lakes also includes laminated, fine sand, silt and clay of lacustrine origin in the Ashuelot valley at Gilsum.

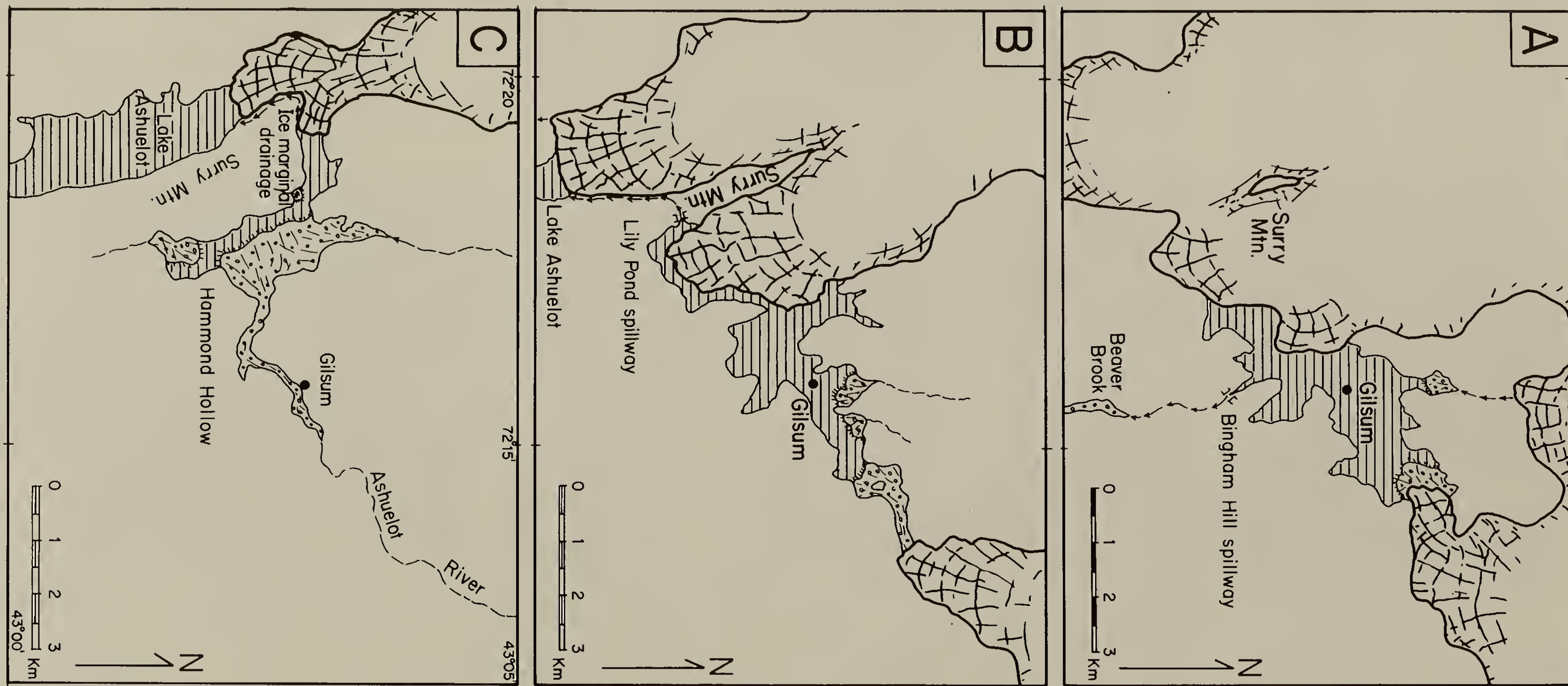


**Figure 6** - Ice deployment during deglaciation of the upper Ashuelot River valley near Gilsum and the impoundment of glacial lakes.

**A. Stage I** - Glacial lake that drains over a col at Brigham Hill (Stop 4, 345 meters) to Beaver Brook is dammed by ice in Hammond Hollow. Ice-contact delta deposited east of Gilsum (Stop 3).

**B. Stage II** - Glacial lake drains over col on Surry Mountain at Lily Pond (320 meters) into the lower Ashuelot valley. Ice-contact delta east of Gilsum (Stop 3) is downcut due to drop in base level. Fine-grained lacustrine sediments are deposited in Ashuelot valley near Gilsum.

**C. Stage III** - Ice recession allows either proglacial or subglacial drainage of lakes in the upper Ashuelot valley around the north end of Surry Mountain (200-260 meters).

Ice wastage in the upper Ashuelot valley appears to follow the overall trend of lobate ice recession in upland valleys adjacent to the Connecticut River. A northeast-southwest trending ice margin in the Ashuelot region is demonstrated by ice persistence in the Ashuelot valley near Surry that is required for the damming of lakes in ice-free portions of the Ashuelot valley to the east. At least part of the Ashuelot valley was ice-free when laminated lacustrine sediments were deposited at Gilsum. Ice appears to have receded from the Ashuelot valley east of Gilsum prior to the lowering of lake levels below 300 meters. Rapid recession and stagnation of ice in the Ashuelot valley, and in other upland tributaries adjacent to the Connecticut Valley, may have been facilitated by the restriction of ice flow across uplands to the north as ice thinning occurred.

## LAKE ASHUELOT

During and persisting until after ice recession in the Ashuelot valley, a large glacial lake, Lake Ashuelot, occuppied the valley from south of Keene to north of Surry (Figure 1). This lake is recorded by fine-grained lacustrine sediments in Keene and many ice-contact and inwash deltas (Fred Larson, pers. com.). The exact position or cause of damming that formed Lake Ashuelot is not known. Initially, damming by ice-marginal deposits south of Keene in the Ashuelot Valley or by a Connecticut Valley lobe may have occurred.

Dissected remnants of a large delta that extended from the north end of Surry Mountain to south of Surry (5 kilometers), are the product of river drainage into Lake Ashuelot (Figure 7). The Surry delta has progressively coarser sediment to the north where it is continuous with a bouldery fluvial terrace. The Surry delta shows no evidence of ice-contact and appears to be sourced by not only Ashuelot River drainage but erosion of till in side tributaries. Exposures in the Surry delta, that show subaerial gravels over subaqueous fine-grained sediment are believed to be the result of delta progradation into a rapidly-infilling, shallow, stable body of water at an elevation of about 170 meters near Surry. Quiet water clayey silt and fine sand deposits that are found directly beneath the fluvial (topset) beds of the delta are believed to be shallow water facies of a gently sloping delta front. The size of the Surry delta, and thus the relatively long time needed to construct the delta, argue for progradation mostly after the disappearance of stagnant ice north of Surry. The sediment in the Surry delta is probably derived primarily from erosion of till and deltaic deposits in the Ashuelot River valley. Evidence for local sources rather than direct glacial sources are large alluvial fans that are graded to the level of Lake Ashuelot and are coeval with the Surry delta. Many of these fans eminate from tributary valleys on the west side of the Surry basin where deeply-incised till deposits occur.

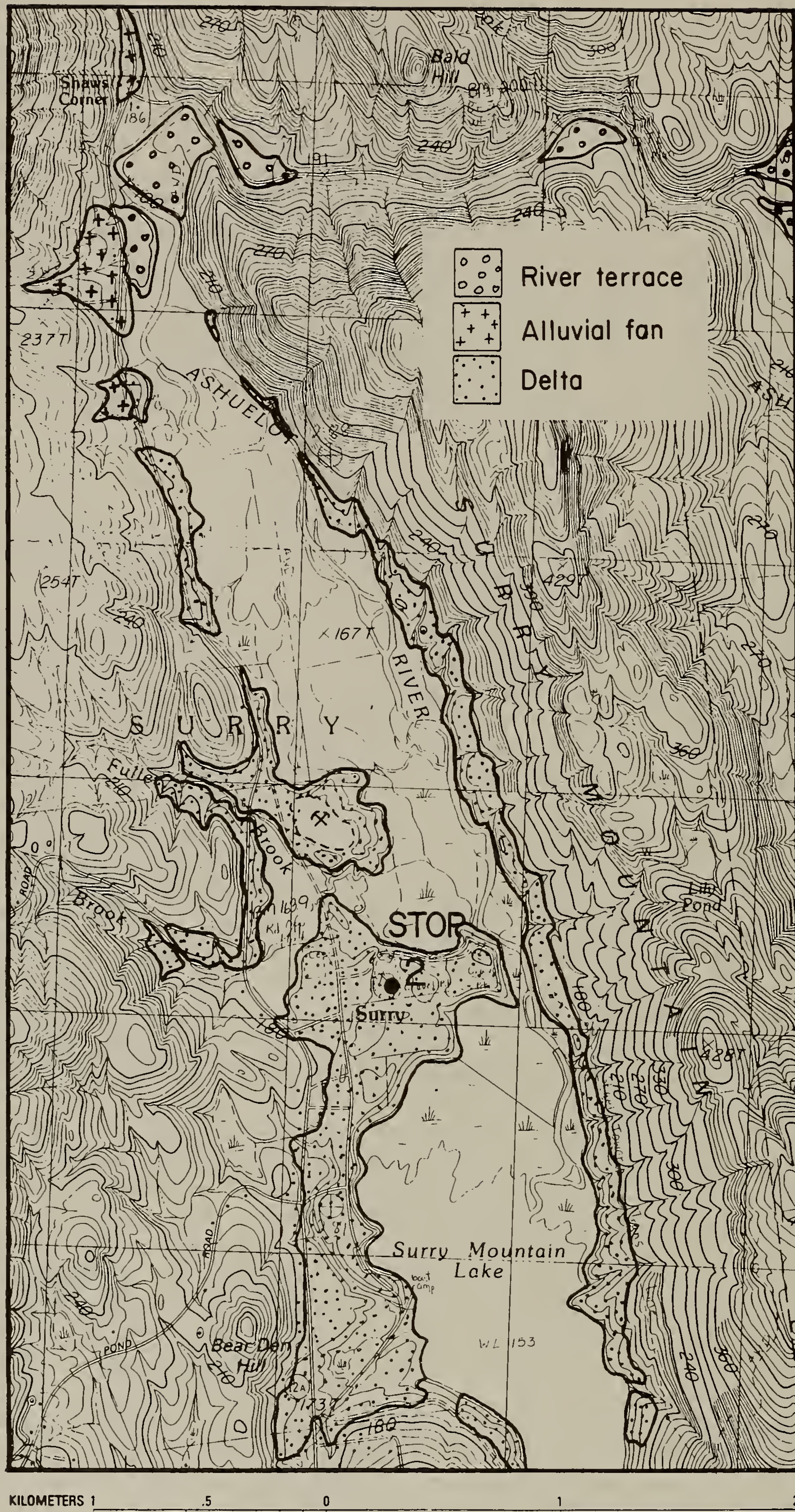


**Figure 7** - Remanent deposits of the Surry delta (Stop 2) and Ashuelot River terrace which are graded to Lake Ashuelot. Also shown are alluvial fans that were deposited by side tributaries and are coeval with the Surry delta. Base is Walpole (1:25,000) Quadrangle. Contour interval: 6 meters.

## ICE RECESSION IN THE WARREN BROOK AND LOWER COLD RIVER VALLEYS

Ice-contact deposits of the lower Cold River valley were formed along the southeast valley wall at a northeast-southwest trending margin of a receding ice lobe that filled the Connecticut valley. Fine-grained lacustrine sediments and deltas in the vicinity of South Acworth (Figure 8) provide evidence of damming by persistent ice to the west. In the Cold River Valley recession of ice from the south side of the valley created a series of ice-marginal ponds and spillway channels that were not developed on the north side of the valley (Figure 8). Meltwater channels (Stop 7), complete with potholes, and deltaic terraces displaying evidence of ice-contact (Stops 5 and 9), are abundant on the south side of the valley from South Acworth to Drewsville. West of Drewsville there is scanty evidence of ice-contact deposition into Lake Hitchcock which extended up the Cold River valley to Alstead and had a water plane at 150 meters (Koteff and Larson, in press). A pausity of ice-contact Lake Hitchcock deltas west of Drewsville could be the result of the planing-off of ice-contact deposits to a lower elevation by later drainage in the valley or the possibility that few ice-contact deposits were constructed.

## LAKE HITCHCOCK - BASE LEVEL CONTROL IN THE COLD RIVER VALLEY

Glacial Lake Hitchcock, which formed in the wake of the Laurentide ice sheet in the Connecticut valley as it receded from Connecticut to northern New Hampshire and Vermont, was the dominant post-glacial base level control in the Cold River valley. Deltas at the mouth of the Cold River extended well down the Connecticut valley and merged with deltas deposited out of the Saxtons River valley in Vermont (Figure 9). Ice-contact deltaic deposits near Drewsville have topset/foreset contacts of at least 150 meters, while most deltaic deposits down valley have topset/foreset contacts at elevations of about 137-139 meters (Koteff and Larson, in press; Carl Koteff, pers. com.). Delta tops in the lower Cold River valley are graded to lower levels of Lake Hitchcock that probably resulted from dam erosion at Rocky Hill, Connecticut and partial lake drainage.

Lake Hitchcock deltas with topset/foreset contacts at lower elevations may be the result of either progradation of deltas into a lower water body, or degradation (scalping) of originally higher deltas by streams graded to lower water levels in the Connecticut Valley. The second mechanism does not record delta progradation but the modification of existing higher delta surfaces, possibly built originally as ice-contact deltas. Existing evidence in the Cold River valley argues for delta progradation into a lowered level of Lake Hitchcock rather than scalping of existing high level deltas. Evidence for delta progradation are listed below.

1. Delta facies suggest topset/foreset deposition with a minimal amount of foreset scalping. The upper parts of foreset packages have a shallow water facies that would be

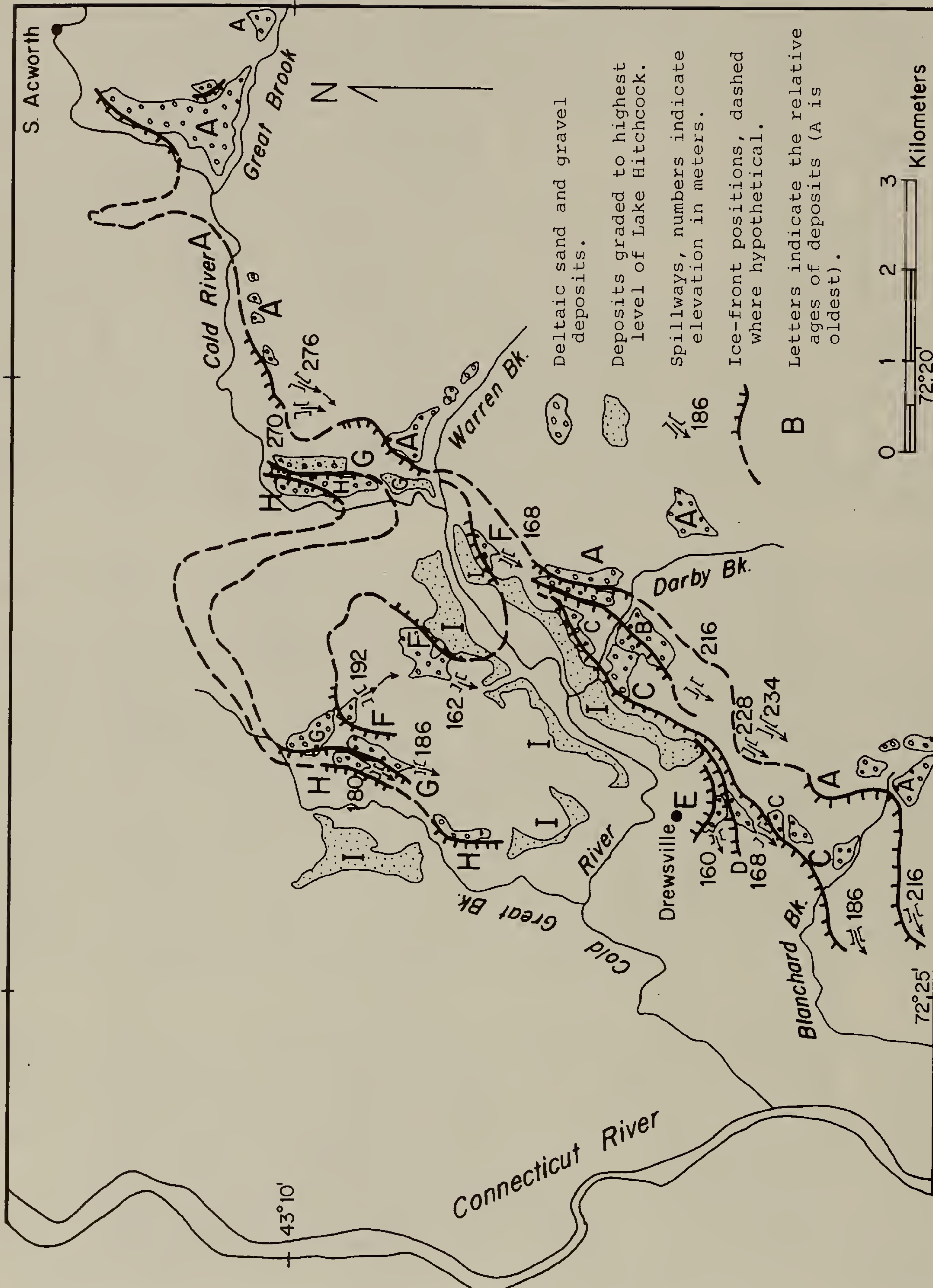


**Figure 8** - Ice-marginal deposits and drainage channels of the Cold River valley from South Acworth to the Connecticut River.

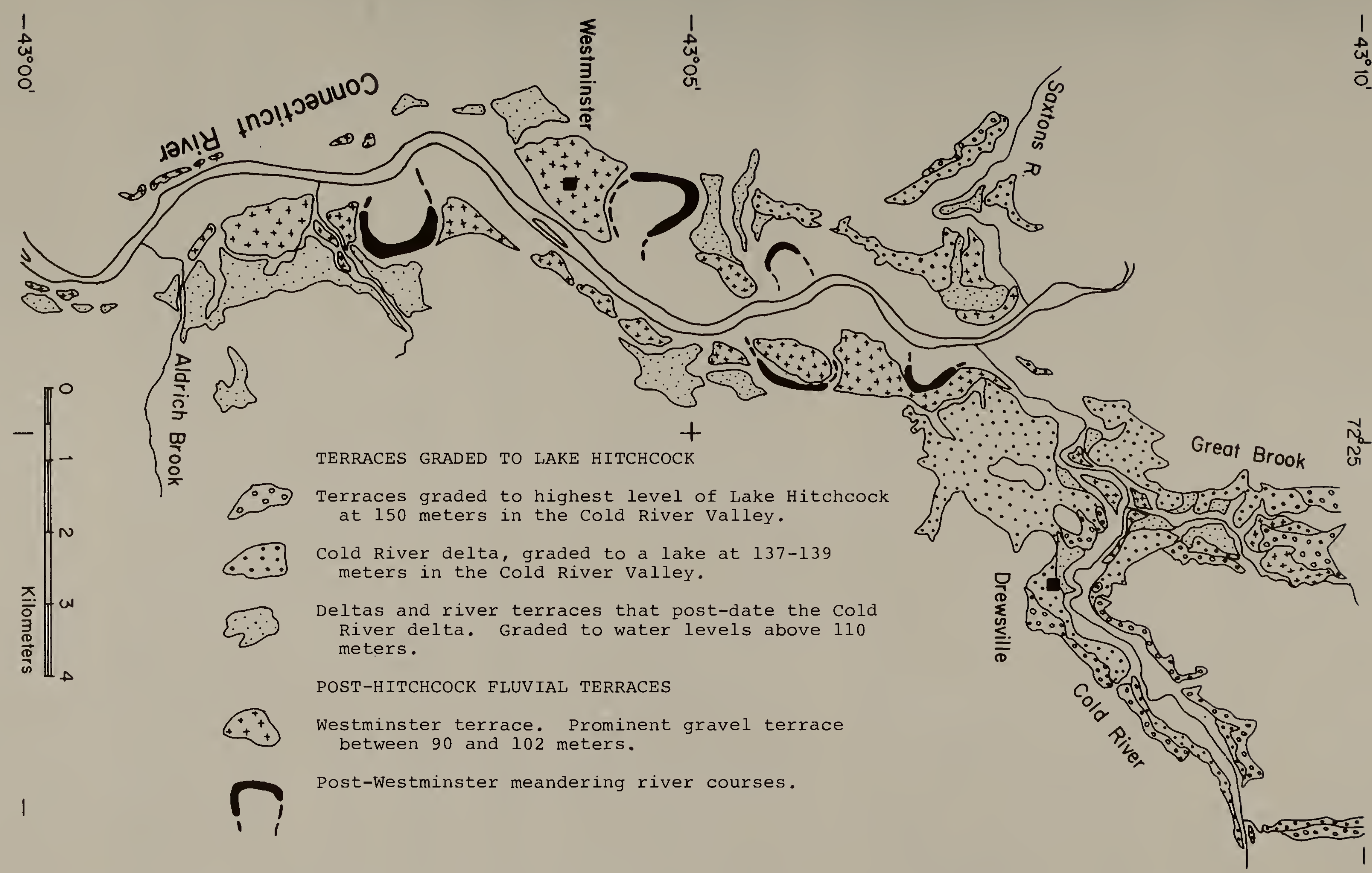


**Figure 9** - Terraces in the Connecticut River valley and its tributaries which are either graded to Lake Hitchcock or were formed by later fluvial systems.

removed by foreset scalping and some foreset beds are continuous with slipfaces on topset channel bars.

2. No remnant delta tops, graded to Lake Hitchcock at 150 meters, have been found in the Cold River valley west of Drewsville in a large area covered by delta tops that are graded to a 139-meter level of Lake Hitchcock. The surfaces of lower level deltas are flat and, except along the Cold River, bear no evidence of inset development or dissection in response to erosion by streams graded to lower lake levels.

3. Lower level deltas have 5 to 8 meter thick topset beds that require substantial delta progradation to accumulate. If deltas were scalped by later river drainage, subsequent topset aggradation could only occur if progradation at the delta front continued after (not during) foreset scalping. Topset beds of 8-meter thickness would require delta progradation of at least 2.5 kilometers or more if a delta slope of less than 3.25 m/km or 2.46 m/km prior to isostatic tilting (17.2 ft/mile, 13 ft/mile prior to tilting) is assumed. Therefore, significant delta progradation occurred in the Connecticut valley after water levels of Lake Hitchcock fell to 139 meters.

4. Varved silt and clay beds beneath the Cold River delta are thick enough (at least 20-25 meters) to require a long period of deposition. Given what appears to be fairly rapid ice recession, it is unlikely that ice-contact deltas would remain active long enough to prograde over thick accumulations of varved sediment. Therefore, it appears that the Cold River delta prograded at some time after the construction of ice-contact deltas in the Cold River valley.

## DRAINAGE OF LAKE HITCHCOCK AND CONNECTICUT VALLEY STREAM TERRACES

A detailed history of how Lake Hitchcock drained in the upper Connecticut valley has not been formulated. Continuous stable water levels for the whole Lake Hitchcock basin, below that into which ice-contact deltas were deposited, have not been recognized. In the Cold River valley progradation of ice-contact deltas into Lake Hitchcock occurred at an elevation of about 150 meters (Figures 8 and 9). A large, flat-topped delta extending for about 4 kilometers down the Cold River and Connecticut valleys from Drewsville, appears to have been built into a stable water level at about 137-139 meters. Lowering of Hitchcock water levels to 139 meters may be the result of initial downcutting of the Lake Hitchcock dam at Rocky Hill, Connecticut.

The lowest Lake Hitchcock water levels, recorded by topset/foreset contacts in the vicinity of the Cold River, occur at the mouth of Aldrich Brook, approximately 10 km to the south (Figure 9). Topset/foreset contacts at this site are at about 120 meters and appear to slope down about 1-2 meters towards the

outer margins of the delta indicating delta progradation into a rapidly falling water body. When one accounts for post-glacial isostatic tilting, these water levels would correspond to an elevation of about 128 meters in the Cold River valley to the north. River terraces, inset in the Cold River delta along the Cold River and Great Brook, appear to be graded to this low level lake (Figure 9).

In the Connecticut valley, from the Cold River to Putney (14 km) a series of gravel terraces occur at elevations of between 90 to 102 meters (Westminster terrace). The terraces are generally composed of truncated fine-grained lacustrine sediment, or the coarser sediments of degraded deltas of Lake Hitchcock, that are capped by 1 to 4 meters of fluvial sand to cobble gravel. The fluvial sediments are coarsest near the mouths of major streams and appear to be derived from the dissection of older glacial and Lake Hitchcock deposits. The terraces were constructed by downcutting, after what appears to be the drainage of lakes in the upper Connecticut valley, and later deposition by the first stable river system to occupy the valley. The Westminster terrace may have been graded to a temporary water body downstream in the Connecticut valley or possibly a bedrock threshold at Lily Pond near Turners Falls, Massachusetts.

Terraces below the 90-meter elevation are generally composed of lacustrine sediments with a cap of 1 to 5 (or more) meters of fluvial medium to fine sand and silt. Terraces below about 78 meters are submerged by the highest modern floods and have deposits capping them associated with those events and eolian activity. Cutoff meanders and oxbow lakes coeval with a terrace level below 78 meters, and that occur at least 6 meters above the modern level of the Connecticut River, indicate that the river had a sinuous channel at some time in the past. The modern Connecticut River channel occurs at a lower level and has a much straighter channel. The meandering channel system may represent the establishment of the Connecticut River after drainage of Lake Hitchcock but prior to complete post-glacial isostatic tilting. Subsequent isostatic tilting may be responsible for straightening the channel to its present course during the earliest Holocene or latest Pleistocene.

## ACKNOWLEDGEMENTS

The author is particularly indebted to Gail Ashley, Carl Koteff, Fred Larson, Phil Shafer, Byron Stone, and Janet Stone for their thoughtful and provacative discussions related to many aspects of the Quaternary geology of the Connecticut valley region. Thanks also go to Gene Boudette, N.H. State Geologist, for continued field support necessary for detailed quadrangle mapping.

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## FIELD TRIP ITINERARY

Assembly point is Keene State College at 8:30 A.M. PLEASE BRING LUNCH WITH YOU BECAUSE THE TRIP WILL NOT BE CLOSE TO A CONVENIENT PLACE TO BUY LUNCH AT LUNCH TIME. It will be important to consolidate vehicles as much as possible in that parking will be a problem at some stops. 7.5 x 15-minute (1:25,000-scale, metric) topographic maps: Bellows Falls (VT), Keene, Marlborough, Stoddard, and Walpole Quadrangles. 15-minute (1:62,500-scale) topographic maps: Bellows Falls, Keene, Lovewell Mountain, and Monadnock Quadrangles.

| Mileage (cumulative) | Trip directions |
|---|---|
| 0.0 | From the Keene State College meeting point proceed to Main Street and head north toward the center of Keene. |
| 0.1 | Turn right (east) onto Water Street. Travel through town and up hill to park where Water Street becomes Peg Shop Road. |
| 1.0 | Stay to right on Peg Shop Road and continue up hill past water tower. Pass Roxbury Street on left. |
| 1.3 | Turn right (east) onto Roxbury Road. |
| 2.1 | Continue down east side of hill and travel under power lines. Turn right onto dirt road shortly after crossing under power lines. |

STOP 1: Gravel pit beneath power lines, Roxbury Road diamicton exposure. Three superposed units (A thru C on Figure below) are exposed in an area of hummocky topography. Unit C is interpreted to be a supraglacial or proglacial stratified deposit with lenses and beds of diamicton facies 1 (Table 1). Unit B is a sandy diamicton (facies 2) of uncertain origin with a subhorizontal platey structure (unloading fissility?). The basal 0.3 m of Unit B contains pieces or lenses of diamicton like Unit A (facies 6). Unit A is interpreted to be lodgement till. Some critical questions to be evaluated are:

1. Is the upper contact of Unit A indicative of shearing by the ice that deposited Unit B, or some other process related to a change in the character of till deposition?

2. Is the platey structure of Unit B due to shearing or is it related to dewatering, subglacial meltout, or unloading?

3. Is the oxidation of Unit A recent or from prior to the deposition of overlying units?

4. Could all three units be the product of a single glaciation or is there evidence of more than one ice advance?

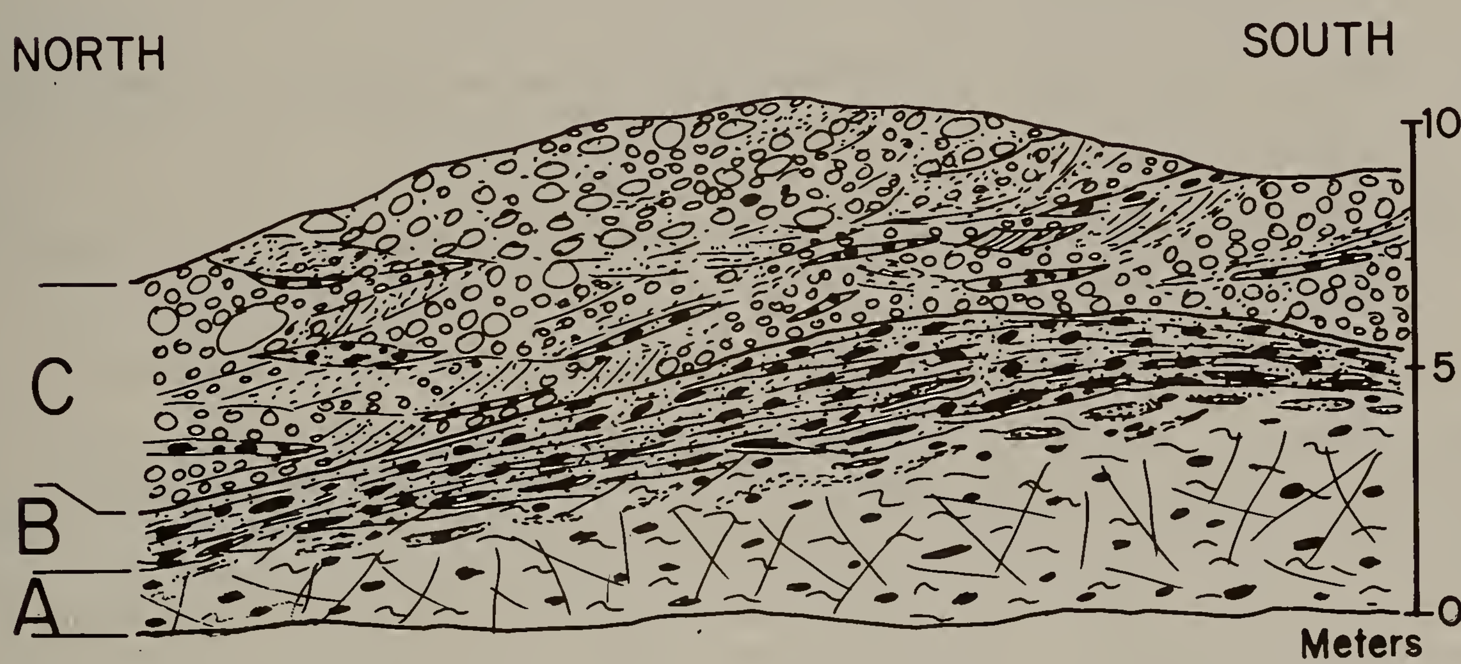


**Figure for STOP 1** - Diamicton stratigraphy exposed on Roxbury Road.

2.9 Return to vehicles and Roxbury Road and head west. Turn left (southwest) onto Peg Hill Road.

3.25 Just beyond water tower (left), turn right onto Roxbury Street.

3.55 After bearing left at bottom of hill turn right (north) onto South Lincoln Street.

3.6 Turn left (west) onto Beaver Street. Cross over top of delta deposited by drainage into Lake Ashuelot from the

Beaver Brook valley.

3.9 Cross over Beaver Brook.

4.0 Continue on Beaver Street to Washington Street and turn right (north).

5.4 Continue north on Washington Street to Rt. 9. Turn left onto four lane highway (Rt. 9).

6.6 Cross over Ashuelot River. Bear right (north) 100 meters beyond river onto Rt. 12 toward Surry.

7.9 Exit right (northeast) onto Maple Ave. towards Surry. Proceed northeast to Rt. 12A.

8.6 At traffic light continue across intersection and join Rt. 12A toward Surry (north).

10.4 Rt. 12A travels across flat tops of deltas built into Lake Ashuelot. Pass entrance to Surry Mountain Dam.

11.5 Pass entrance to Surry Mountain Reservoir campground and recreation area.

12.6 Pass Crain Road which leads to town of Surry (village to east) that sits on top of Surry delta that was built into Lake Ashuelot.

12.9 Turn right (east) onto dirt road that leads to gravel pits in Surry delta.

13.3 Follow dirt road to gravel pits east of town of Surry.

STOP 2: Gravel pits in Surry delta (see Figure 7). Gravel pit exposures reveal the distal portion of the delta which was deposited by the Ashuelot River and the side tributaries of the Surry valley. Subaqueous deposits of the Surry delta are continuous beds of laminated and rippled clayey silt to medium sand that achieve an elevation of about 170 meters. Subaerial (topset) beds are mostly medium to coarse, trough-crossbedded sand with some pebble gravel. Do the sediment facies in the Surry delta indicate delta progradation into a shallow body of water or the trimming of a delta to a lower water level and subsequent fluvial deposition after downcutting?

13.7 Return to Rt. 12A from gravel pits and head north.

15.8 Pass by foot bridge over Ashuelot River. Terrace across river is a bouldery fluvial terrace graded to the surface of the Surry delta down valley.

16.0 Turn right (east) at Shaws Corner onto Gilsum Road.

16.2 Terrace across the Ashuelot River at this point is the bouldery gravel river terrace graded to the Surry delta.

16.4 Road follows the coarse of the Ashuelot River. Bedrock is exposed on the south valley wall while thick till benches crop out along the north side of the Ashuelot River. Alluvial fan on the north side of the road at this point is continuous with the surface of the terrace that is graded to the Surry delta and the fan is formed primarily from dissected till of benches to the north.

17.3 Pass the mouth of Cannon Brook. Thick till benches occur in this area in the lee of Bald Hill.

17.8 Road departs Ashuelot River and parallels Dart Brook which drains the upland to the north. The Ashuelot River crosses the basin of Hammond Hollow that is filled with deltas deposited in a lake dammed by ice to the west during ice recession. The south end of Hammond Hollow is filled with ice stagnation deposits and a sandy late Wisconsinan diamicton. River terraces in the valley are graded to the Surry delta.

18.2 Turn right at Roundys Corner.

19.0 Road crosses the surface of the highest delta in the Hammond Hollow basin. VERY SHARP left turn descends to the Ashuelot River through thick till of a till bench.

19.2 Pass bridge to Hammond Hollow.

19.9 Steep bank on left exposes till in benches that occur on the lee side of the bedrock upland to the north. Below road to the right the Ashuelot River passes over a series of rapids.

20.0 Pass over the Ashuelot River. Rapids below to the right (west). Turn left onto Rt. 10 toward Gilsum.

20.4 Laminated silt and clay are exposed along west bank of Ashuelot River at north end of bridge.

20.6 Town of Gilsum. Pass by road that leads through center of town. Laminated silt and clay are exposed along the east bank of the Ashuelot River at this point.

21.0 Cross over Ashuelot River again.

22.0 Pull into gravel pit on north side (left) of Rt. 10.

STOP 3: Gravel pits in deltas east of Gilsum. Exposures reveal the interior of a large ice-contact deposit that prograded from ice to the east in the Ashuelot Valley. Steeply dipping pebble and cobble gravel and coarse sand foreset beds on the north side of Rt. 10 show evidence of ice-contact in the form of collapse faults and were formed either on a prograding delta front or a subaqueous fan. On the south side of Rt. 10, fine to medium sand bottomset beds are truncated along an unconformity overlain by

fluvial pebble gravel. Deposition of coarse foresets was probably into a lake that drained at the spillway at Bingham Hill (Stop 4; Figure 6A) or some higher spillway. Later downcutting of subaqueous beds and deposition of fluvial gravels probably occurred when ice recession to the west allowed Ashuelot valley lakes to drain over the col at Lily Pond on Surry Mountain (Figure 6B).

24.1 Back track on Rt. 10 through Gilsum to the point where the field trip first encountered Rt. 10 southwest of Gilsum. Continue on Rt. 10 (south).

25.1 Follow Rt. 10 south to top of hill. Pull into parking area just south of divide.

STOP 4: Pot Holes and Bears Den State Forest, Bingham Hill. Follow the trail east from the parking lot to a col on Bingham Hill that has many potholes formed by drainage down the east flank of the hill. The potholes are not related to lake drainage but mark the place where meltwater issued from the margin of a receding ice mass or flowed subglacially. The steep crest of Bingham Hill may have been the site of crevassing or nunatak formation that would have allowed the drainage of supraglacial meltwater to subglacial cavities during ice recession.

Continue down the east side of Bingham Hill. The ravine 0.1 miles east of Bingham Hill, that is now floored by a swamp, was a spillway channel that drained a lake in the Ashuelot valley at an elevation of about 345 meters. The small cliff face at the north end of the swamp is a small cataract formed by the discharge of water across the spillway divide to the north.

26.8 Return to Rt. 10 and head back toward Gilsum (north). At the town of Gilsum turn left (north) toward center of town.

26.9 Pass by the monument in the center of Gilsum that is decorated with pegmatite minerals including some large green beryl crystals. The uplands surrounding Gilsum are known to mineral collectors for their pegmatite mines which have been inactive since the 1950's. Gilsum is also the site of the "Rock Swap" which is an annual meeting of mineral collectors. If you need to purchase food for lunch, the Gilsum store across the street from the monument will be your last opportunity. The caravan will continue north to the lunch stop at Kidders Pond which is an abandoned pegmatite mine.

27.5 Continue north on road through Gilsum. Pass rotten schist outcrop on the left (west) near top of hill.

28.2 Pass large swamp on left. The upland area north of Gilsum is underlain by pegmatite-bearing schist and is almost entirely barren of glacial sediment. Striations are not preserved on the schist but occur on almost every outcrop of pegmatite which usually contains large quartz grains that preserve striations.

28.8 Pass old mine shaft on left (west). Most dirt roads in the area lead to pegmatite mines in the adjacent hills.

29.4 Just before sharp bend to left turn right (east) onto dirt road (Easy to miss!) made of white feldspar fragments that leads to an old tipple on Kidders Pond.

LUNCH STOP: Head 0.4 miles east to east side of Kidders Pond. Gem quality aquamarine crystals have been found at the southwest corner of the pond. After lunch go back to main road and turn right (north) towards Lake Warren.

30.7 On the right (east) is schist that is heavily striated. On the left (west) is a meltwater channel that allowed water in the Warren Brook drainage basin to escape to the south into Dart Brook and the Ashuelot River drainage system (Figure 6C). Deltas built from the mouth of Dart Brook in the Ashuelot valley are small and composed of mostly pebbly sand. For this reason it is thought that glacial drainage from the Warren Brook valley was not a major source of sediment that was deposited in Ashuelot valley lakes. The divide between Dart Brook and Warren Brook drainages is at the southern end of the trailer park to the north which sits on a sand and gravel deposit formed behind the divide.

32.5 Turn left (west) at intersection with Rt. 123 at town of East Alstead. Overlook of Lake Warren to the west.

33.2 Pass Old Settlers Road which comes into Rt. 123 from the right (north).

33.6 Cross over Warren Brook at old mill. Continue west on Rt.123.

35.3 Cross over Forristal Road at north end of swamp. Diamicton exposure to west on Forristal Road is a very sparsely stony, silty till (facies 5) that appears to be composed of reworked lake sediment and is in the base of a till bench. Rt. 123 follows Warren Brook from this point on and along the road are many exposures of thick diamicton.

36.6 Rt. 123 merges with Rt. 12A which comes in from the left (south).

37.1 Pull over to right (north side) of Rt. 123-12A at large bluff exposure on Warren Brook.

STOP 5: Large bluff section on Warren Brook (see figure on following page). Along Warren Brook is an ice-contact delta and lake bottom sediment which rest on subaqueous diamicton (facies 3) and unoxidized, gray, late Wisconsinan till (facies 4). At stream level, below slumped material at the western end of the outcrop area, the Warren Brook beds are sometimes exposed. The upper contact of the Warren Brook beds is not exposed at this stop but it will be seen at Stop 6. Early post-glacial erosion in the Warren Brook valley is evident from the till surface that

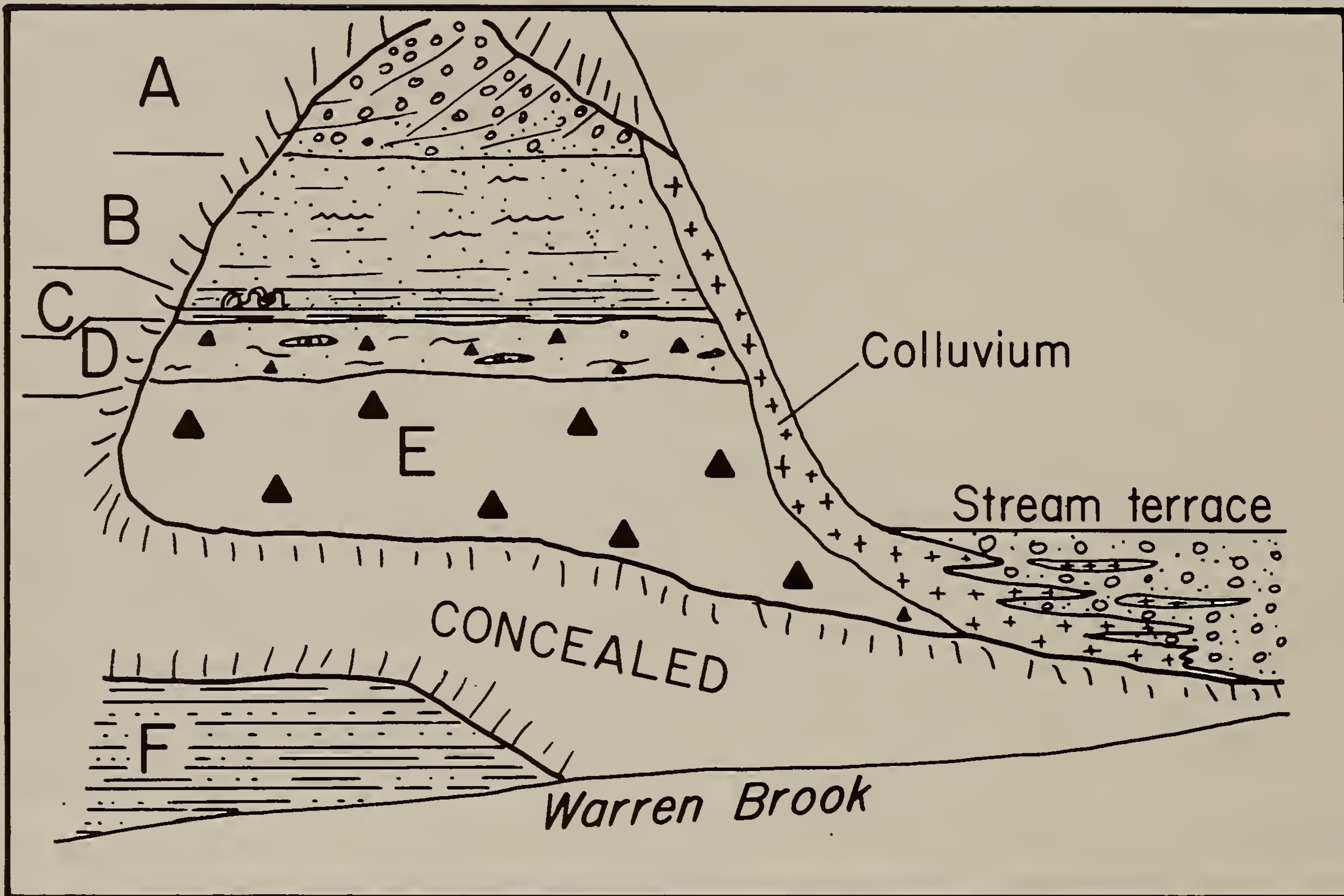


**Figure for Stop 5** - Large bluff section on Warren Brook.

Unit A — 10 meters, bouldery to pebbly, steeply-dipping gravel beds. Deltaic foreset beds.

Unit B — 12.5 meters, fining downwards from massive flat beds of coarse to medium sand to laminated and rippled, fine to medium sand with occasional clayey silt beds. Deltaic bottomset beds.

Unit C — 3 meters, laminated fine sand to dark gray silty clay beds. Soft sediment deformation. Lake botom sediments.

Unit D — 3.5 meters, crudely layered dark gray diamicton (facies 3) with lenses of sand and clayey silt and pebble concentrations. Load deformation. Subaqueous sediment flow or pelagic rainout deposits.

Unit E — 12-14 meters, compact, fissile, dark gray silty diamicton (facies 4). Lodgement till.

--- about 10 meters concealed ---

Unit F — 7 meters, laminated, medium to light gray fine sand and silt. Warren Brook beds. Lake bottom sediments.

was downcut prior to the aggradation of bouldery gravel that forms a terrace (east end of outcrop) graded to deltas of Lake Hitchcock.

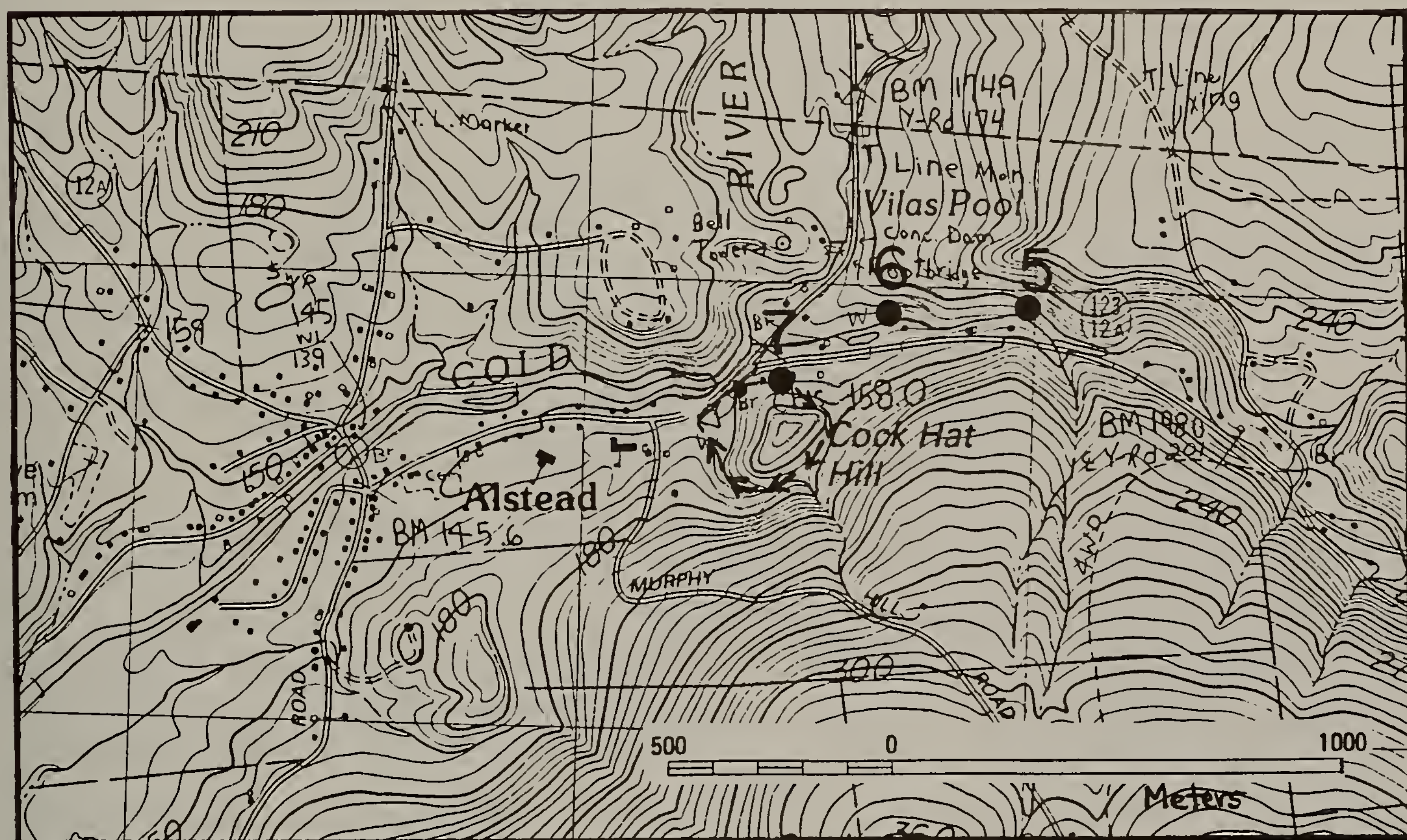


**Figure for Stops 5 thru 7** - Location of stops. Base is Bellows Falls (1:25,000) Quad.

37.4 Return to vehicles and proceed west to the junction of Rt.123A. Turn right (north) onto Rt. 123A. Pass over Warren Brook and immediately turn right (east) up the small dirt road leading into pasture along north bank of Warren Brook.

<u>STOP 6</u>: Small bluff on Warren Brook (see figure on following page). Beneath bouldery gravel, that caps the terrace-like surface north of Warren Brook, is till (facies 5) overlying silt and sand of the Warren Brook beds. The top of the Warren Brook beds are deformed by overriding ice and the till contains sheared lake sediment and deformed weathered bedrock. The Warren Brook beds drape over colluviated schist which in turn lies over an outcrop of weathered, non-colluviated schist. The exposure records the proglacial damming of a lake in the Warren Brook valley as a glacier advanced southward in the Connecticut valley.

WALK Do not reboard the vehicles. Walk back out to Rt. 123A and across Warren Brook to the south side of Rt. 123. Congregate north of Cock Hat Hill on Rt. 123. Follow trail through woods onto terrace at the beginning of a meltwater channel east of Cock Hat Hill.

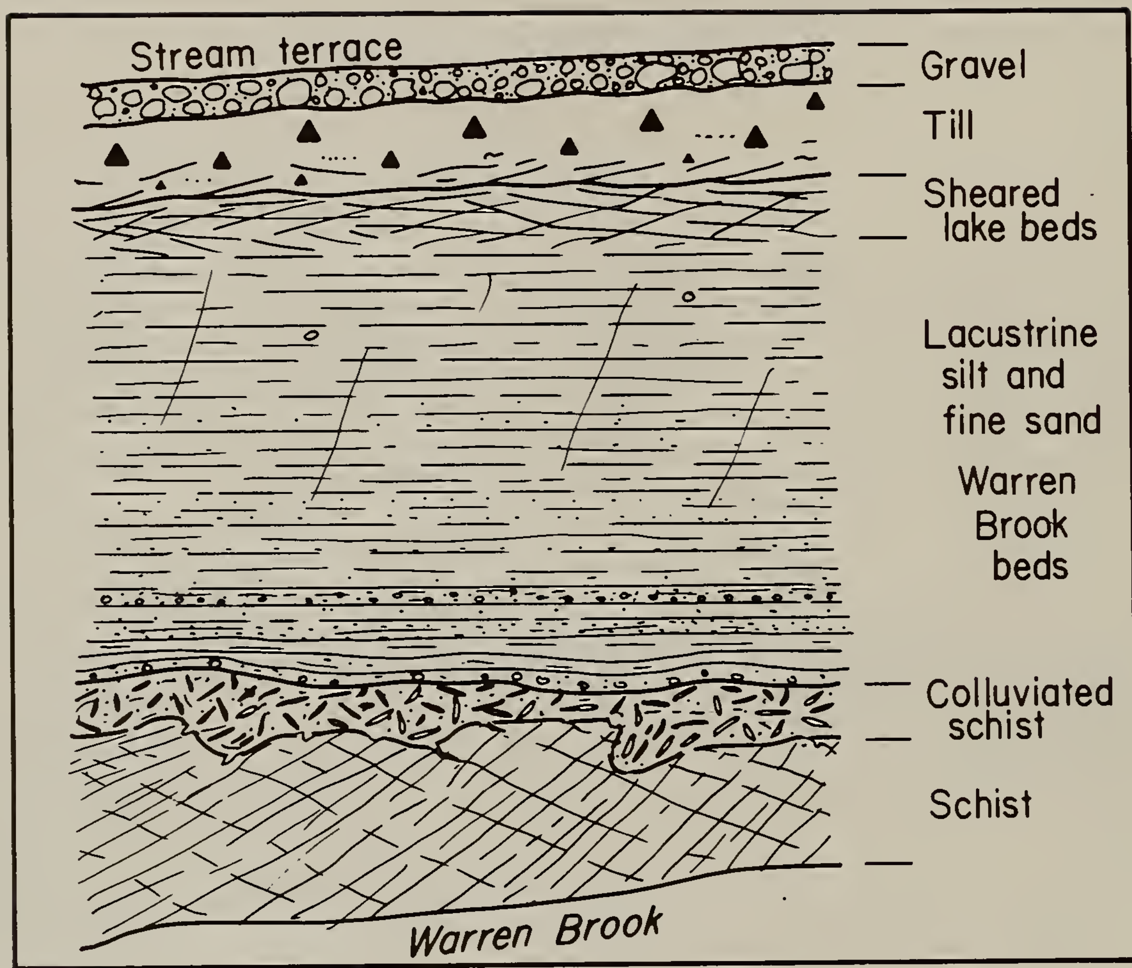


**Figure for Stop 6** - Small bluff section on Warren Brook.

STOP 7: Cock Hat Hill meltwater channel and fluvial gravel over saprolite. The terrace at the entrance to the meltwater channel is composed of bouldery gravel and is graded to the deltas of Lake Hitchcock. The north side of Cock Hat Hill has 15-meter exposures of till (facies 5) which is actively slumping onto the terrace below. Walk through the meltwater channel which swings south of Cock Hat Hill and enters the Cold River valley again at a series of potholes cut in gneiss. The channel was probably cut as meltwater (subglacial or ice-marginal?) made its way along the south valley wall of the Cold River valley. The channel was abandoned upon ice recession and reactivated as terraces aggraded behind Lake Hitchcock deltas reached the elevation of the channel mouth. The north face of Cock Hat Hill exposes about 20 meters of till down to the channel level but bedrock is exposed on the south side of Cock Hat Hill up to 10 meters above the channel floor. Schist is exposed along the entire valley wall across from Cock Hat Hill and it is suspected that the entire channel floor is underlain by bedrock.

After viewing the potholes west of Cock Hat Hill proceed across the field to Rts. 123-12A and walk east along the highway. CAREFUL! - Traffic is dangerous at this point. Along the highway is an exposure of 4 meters of bouldery gravel over 5-6 meters of schist saprolite. The gravel is a continuation of the terrace at the entrance to the meltwater channel and capping the small bluff exposure on Warren Brook (Stop 6). Meltwater that drained around Cock Hat Hill may have taken advantage of saprolite in cutting the meltwater channel. It is suspected that till, like that exposed on the north flank of Cock Hat Hill, or the Warren Brook

beds were originally overlying the saprolite prior to fluvial downcutting and later terrace aggradation behind Lake Hitchcock deltas.

38.0 Return to vehicles and head back to Rts. 123-12A. Turn right (west) onto Rt. 123 toward the town of Alstead. Center of the town of Alstead. Turn left (south) onto Hill Road.

38.5 Off to the right (west) are several levels of Lake Hitchcock deltas that fill the Cold River valley.

39.3 Road crosses over Darby Brook. Stop 8 is bluff visible on bank dead ahead to the east.

39.6 Continue south on Hill Road to intersection with Mill Road. Make a 180-degree left towards Cook Hill Farm.

39.8 Stop at side of road near house on bluff.

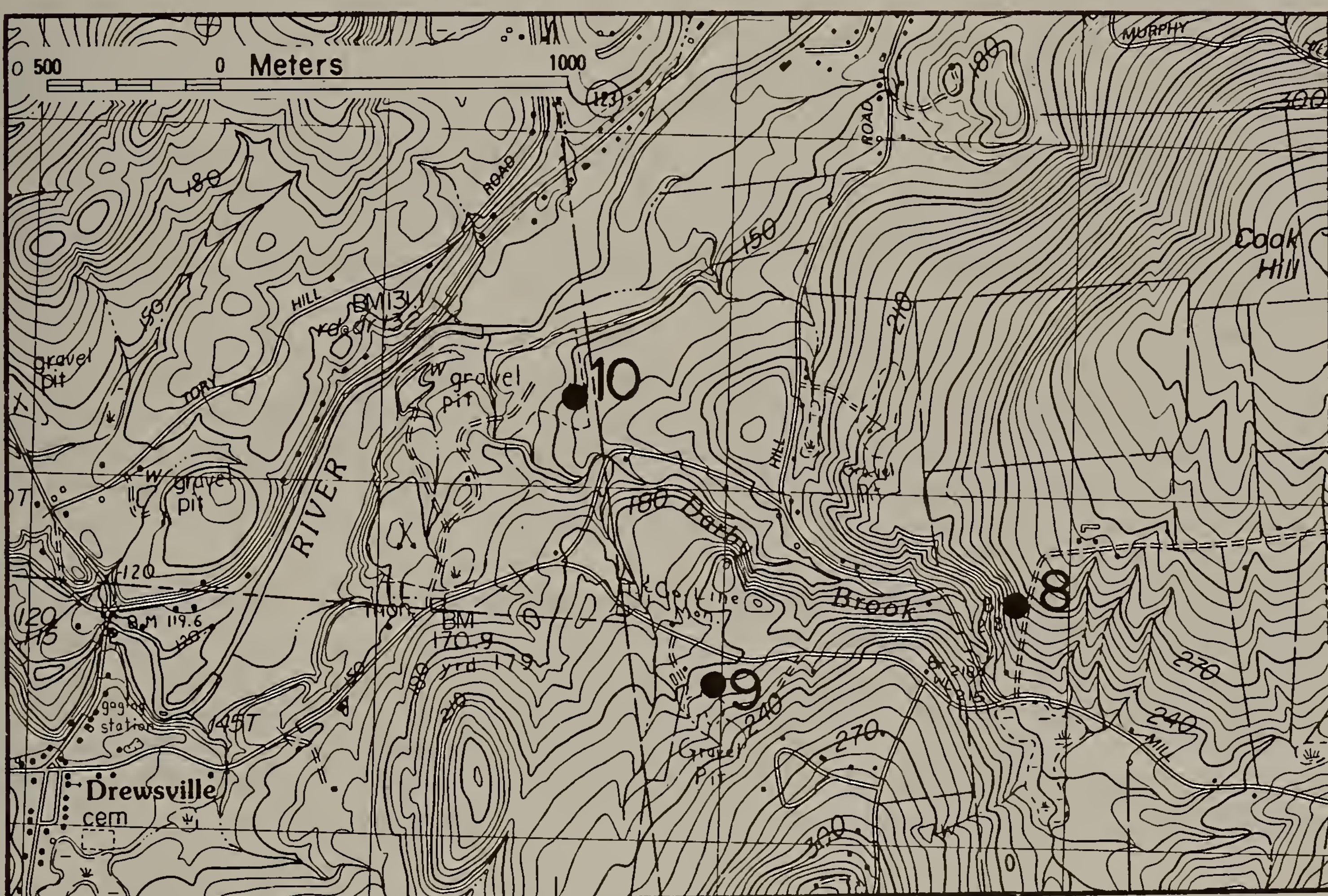


**Figure for Stops 8 thru 10** - Location of stops. Base is Bellows Falls (1:25,000) Quadrangle.

STOP 8: Bluff exposure west of Cook Hill Farm Road. Darby Brook has undercut a till bench formed in the lee of Cook Hill. (Tall tales till tails tell!)

Description of section: (Top to bottom)

| | |
|---|---|
| 20 meters | Dark olive gray (oxidized) pebbly diamicton with a clayey silt matrix (facies 5). Lodgement till. |
| 2-3 meters | Laminated fine to medium sand and silty clay beds that contain abundant dropstones and non-lithified drop sediment ('drop clots'). Intratill stratified deposit (subglacial?). |
| +2 meters (limit of exposure) | Dark olive gray (oxidized) pebbly diamicton with a clayey silt matrix (facies 5). Extremely compact, may be stonier than diamicton at top of section. |

The exposure above is typical of till benches that contain intratill stratified deposits. About 100-200 meters north of Stop 8 and at the same depth below the top of the till bench as at Stop 8, a gully exposure reveals an intratill deposit. At the gully exposure the intratill deposit is a very crudely stratified muddy gravel composed of abundant angular schist blocks derived from the bedrock of Cook Hill and fewer exotic lithologies than occur in the surrounding diamictons. The gravelly pocket of intratill sediment is believed to be a mass movement deposit and an up-slope correlative of the fine-grained stratified deposits at Stop 8.

A water well drilled at the house at Stop 8 penetrated 75 meters of (240 feet) of surficial material before hitting bedrock. Bercause bedrock is exposed along Darby Brook, about 50 meters (160 feet) below the house at Stop 8, it is suspected that the water well penetrated either a buried valley or as much as 25 meters of saprolite beneath the till bench.

40.0 Return to Mill Road at the base of Cook Hill and turn right. Continue west on Mill Road.

40.2 Top of hill on Mill Road. Mill Road sits on a till bench formed on the north side of a hill. View back to the right from top of the hill is of the bluff exposure at Stop 8 and the till bench south of Cook Hill. Continue west.

40.5 Pull over to side of Mill Road at base of hill near large farm. Walk to sand and gravel pits south of road in pasture.

STOP 9: Small gravel pits on Mill Road. Exposed are deltaic and lake-bottom sediments that were deposited in a small proglacial pond trapped against the south side of the Cold River valley (Deposits B on Figure 8). Deposits of this type are common on the south valley wall from Drewsville northeast to South Acworth.

40.8 Return to vehicles and continue west. Turn right at

intersection.

41.1 Cross Darby Brook and turn left into dirt road immediately after bridge. Take left fork (low road) into gravel pit.

STOP 10: Gravel pit in ice-contact Lake Hitchcock delta deposited from an ice margin to the northeast near Alstead (coeval with Deposits F or G on Figure 8). East face of pit exposes about 2 meters of bouldery topset beds over 10-15 meters of cobble gravel to medium sand foreset beds (contact at about 150 meters). A kettle infilling of collapsed fine to medium sand is also exposed beneath the topset beds. The north end of the pit shows a lower level topset-foreset contact formed by the trimming of the ice-contact delta by later drainage graded to lowered water levels of Lake Hitchcock. Extremely bouldery beds near the entrance to the pit are dissected deposits of an older ice-contact deposit that is not related to lake Hitchcock. Bedrock is exposed in the floor of the pit and along Darby Brook to the south.

41.4 Return to Mill Road and turn right (west) towards Drewsville. Intersection is at level of a deltaic terrace deposited in another ice-dammed pond above the level of Lake Hitchcock (Deposit C on Figure 8). Mill Road to the west traverses several levels of terraces graded to different levels of Lake Hitchcock.

42.4 Center of Drewsville. Merge with Rt. 123 and continue west on Rt. 123. Just after merging with Rt. 123, road overlooks terrace surface to right graded to what is believed to be the lowest level of Lake Hitchcock (about 128 meters).

42.9 Rt. 123 crosses onto top of Cold River delta which is graded to a level of lake Hitchcock at about 137-139 meters. Up-valley many ice-contact deposits were trimmed by later drainage to this base level. Continue west on Rt. 123.

44.4 Turn left (south) at T-intersection at bottom of hill after descending from the top of the Hitchcock delta. Road now crosses the top of the gravelly Westminster terrace which was cut by post-Hitchcock drainage in the Connecticut valley.

45.7 Intersection with Rt. 12. Turn left (south) on to Rt. 12 towards Walpole.

46.0 Pass exit on left to Walpole.

47.0 Pass Rt. 123 bridge to Westminster Station, Vt. (Closed for construction.

50.0 Cross over ravine of Great Brook.

51.3 Cross over ravine of Houghton Brook.

51.5 Turn left (east) onto road leading to gravel pit.

STOP 11: Gravel pits in Aldrich Brook delta. The Aldrich Brook delta is an inwash delta built into Lake Hitchcock primarily from the drainage of Aldrich Brook as is indicated by the northwest-dipping foreset beds. Houghton Brook, which occurs on the north side of the delta, apparently contributed little sediment to the delta because its basin has a thin glacial cover and little sediment was available for erosion. The Aldrich Brook delta is unusually coarse for an inwash delta and in some places appears to coarsen outwards. Topset/foreset contacts in the center of the delta are at about 120 meters which corresponds to a level of about 128 meters in the Cold River valley to the north. Because topset/foreset contacts in the outer part of the Aldrich Brook delta slope downward at elevations below 120 meters, the outer portion of the delta probably prograded into a rapidly falling lake. The sudden dissection of coarse deposits along Aldrich Brook, in response to falling water levels of Lake Hitchcock, probably accounts for the outward coarsening of the delta. The topset/foreset contacts of the Aldrich Brook delta are the lowest observed for a post-Lake Hitchcock delta in the Walpole Quad. and suggest the last gasp of the lake. The face of the gravel pit to the west shows fluvial incisions that cut into the foreset beds of the Aldrich Brook delta. The incisions were later infilled with eolian sand and pebbly gully-wash deposits.

51.6 Return to vehicles and head south on Rt. 12. Cross over ravine of Aldrich Brook. Low area to southwest of ravine is underlain by fine-grained lacustrine sediments and does not have a fluvial cap. This surface may be a remnant piece of the floor of Lake Hitchcock when it drained. The fine-grained lacustrine sediments, which are exposed in gullies along the Connecticut River to the west, are overlain by a fine sand and silt deposit that is eolian.

64.1 Continue south on Rt. 12 which leads back to Keene.

END OF FIELD TRIP

# THE ASCUTNEY MOUNTAIN IGNEOUS COMPLEX

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## INTRODUCTION

The Ascutney Mountain igneous complex, located in southeastern Vermont, is a member of the White Mountain Plutonic-Volcanic series (figure 1). Approximately 8 kilometers by 4 kilometers in size, this Cretaceous (122 Ma, Foland et al., 1985) complex consists of three stocks and a crescentic ring dike. Reginald Daly formulated the theory of magmatic stoping as a result of his work at the complex (Daly, 1903). Consequently, Ascutney Mountain is now a classic geologic locality for the study of igneous petrology. The purpose of this field trip is to examine igneous rocks of the complex and the numerous types of xenoliths included in them. In particular, we will focus on xenoliths of breccia in the syenite ring dike of Little Ascutney Mountain. A diverse array of rock fragments in the breccia records the Cretaceous stratigraphy of the region.

## GEOLOGIC SETTING

The Ascutney Mountain igneous complex intrudes the northeast portion of the Chester dome and overlying east-dipping Paleozoic metasediments (figure 2). The Chester dome is comprised of Precambrian basement gneisses that were metamorphosed during the Grenville (~1100 Ma) and subsequent orogenies. The Cambro-Ordovician metasediments flanking the dome were metamorphosed in the Taconic (~485 Ma) and subsequent orogenies; the younger Paleozoic metasediments were first metamorphosed in the Acadian orogeny (~380 Ma), a deformational event responsible for the main folding and metamorphism east of the Green Mountain anticlinorium.

The Northfield, Waits River, and Gile Mountain formations of the Connecticut Valley-Gaspé synclinorium constitute the bulk of metasediments cut by the Ascutney complex; they consist of a metamorphosed eugeosynclinal sequence of sandy limestones, calcareous sandstones, and shales. A thin band of mafic metavolcanics, the Standing Pond amphibolite, forms a canoe-shaped loop within the sequence and marks the boundary, as mapped by Doll et al. (1961), between the Waits River and Gile Mountain formations, units that are otherwise not easily distinguished (Hepburn et al., 1984; Downie, 1982). Metamorphic isograds trend north-south and metamorphic grade decreases from staurolite-kyanite grade in the core of the Chester dome through garnet and biotite grades to chlorite grade along the Connecticut River (Thompson & Norton, 1968). These relations are depicted in figure 3. A metamorphic aureole is well-developed around the syenite stock of Ascutney Mountain and is superimposed on the regional metamorphic isograds (Nielson, 1973). Note that the Ascutney complex is adjacent to the Monroe line (figure 2), the boundary between Vermont sequence (Doll et al., 1961) and New Hampshire sequence (Billings, 1956; Lyons et al., 1983) rocks.

The geologic history of the Ascutney region comprises three stages: (1) at least two episodes of Paleozoic deformation and metamorphism associated with the Taconian and Acadian orogenic events, (2) intrusion of calc-alkaline magmas causing contact metamorphism, dike emplacement, and explosive volcanism, (3) incorporation of country rock, volcanic and breccia xenoliths by the intruding magmas.

## PETROLOGY OF THE IGNEOUS ROCKS

The Ascutney Mountain igneous complex consists of three intrusive rock phases; in order of decreasing age, these are gabbro-diorite, syenite and granite (figure 3). A variety of dikes cut the plutons.

<u>Gabbro-diorite</u>. Termed the "basic stock" by Daly (1903), the gabbro-diorite stock forms the main mass of Little Ascutney Mountain. Daly (1903) identified five varieties of the "basic stock": augite-gabbro, hornblende-biotite-augite gabbro, biotite-hornblende diorite, biotite-augite-hornblende diorite, and orthoclase-microperthite-

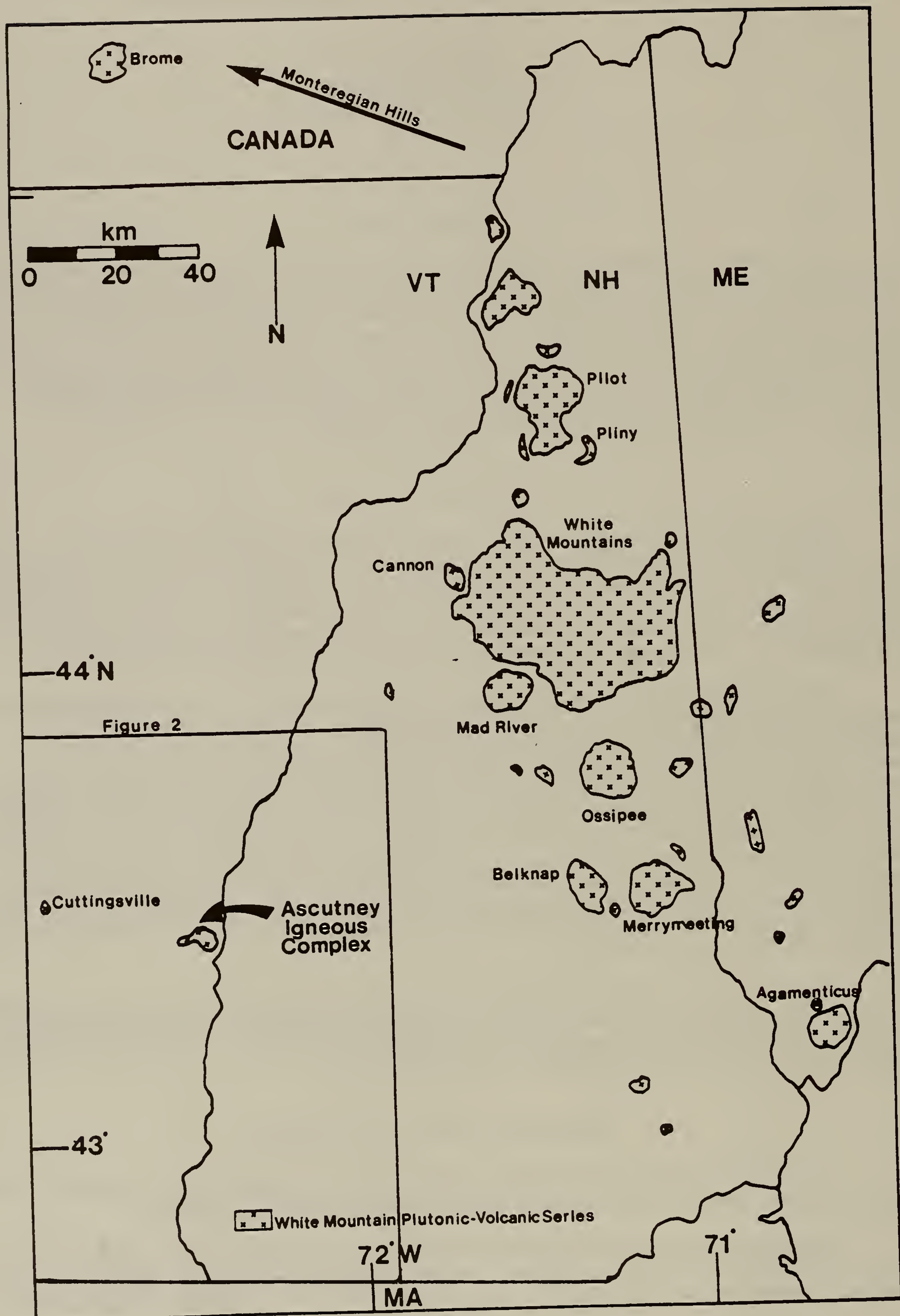


Figure 1. Location map showing Ascutney and other White Mountain plutonic-volcanic series rocks (after Foland & Faul, 1977).

Figure 2. Generalized geologic map of SE Vermont and SW New Hampshire [after Billings,1956; Doll et al.,1961; Downie, 1982].

hornblende-biotite diorite. He also noted the presence of rock types transitional between these. The variety and intimate association of mafic rock types precludes their mapping as individual rock units. In the field the gabbros are distinguished from the diorites by their darker color and, as noted by Chapman and Chapman (1940), the presence of inclusions of the former in the latter indicates that the emplacement of the gabbros preceded that of the diorites.

Rocks of the mafic intrusion are coarse- to medium-grained and have subhedral granular texture. They consist of plagioclase, augite, biotite, hornblende, alkali feldspar and quartz, in varying proportions. Zircon, sphene, apatite, magnetite and ilmenite are the accessory minerals (stop 1-station 7).

Syenite. Syenite constitutes the main mass of Ascutney Mountain, forms a crescentic ring dike on the south side of Little Ascutney Mountain, and intrudes the gabbro-diorite as a ring-shaped plug. Because of the common presence of quartz in the syenites, Daly (1903) called them nordmarkites. He described the main stock as a composite of four rock types: hornblende-biotite nordmarkite, porphyritic hornblende-biotite-augite nordmarkite, alkaline granite and monzonite. As with the gabbro-diorite suite, however, a range of rock types with variant chemistry, mineral assemblages and structure grade into and are closely associated with one another in different locations on the mountain. Therefore, the geographical extent of syenitic rock types is not delineated on the geologic map (figure 3); the entire mass is mapped as syenite (Daly, 1903; Chapman & Chapman, 1940; Nielson, 1973). Likewise, although mapped as syenite, the dike of Little Ascutney is structurally and chemically different from the syenites of the main mountain; it is fine-grained, porphyritic, less Fe-rich than the main mountain syenites, and probably hypabyssal.

The syenites consist of porphyritic, seriate, or subhedral granular medium-grained aggregates of perthite, hornblende and biotite. Minor constitutents are quartz, augite and fayalite with apatite, zircon, sphene, monazite, ilmenite and magnetite as accessories (stops 1,2, and 5).

A variety of xenoliths occurs within the syenite stock. Cognate xenoliths of gabbro and diorite are present and are direct field evidence for the older age of emplacement of the mafic rocks. Other small mafic inclusions with porphyritic structure are very abundant. They consist of perthite, microcline, orthoclase, plagioclase, hornblende and augite phenocrysts in a fine-grained matrix of plagioclase, hornblende, biotite and quartz (stop 2). Volcanic flows and tuffs, first mapped by Chapman and Chapman (1940), occur as a screen in the western portion of the syenite stock and as smaller elliptical bodies north of the northern summit of Ascutney Mountain. Fragments of the country rock through which the magma passed also occur as xenoliths. Large masses of breccia, a fifth type of inclusion, occur in the syenite porphyry dike of Little Ascutney Mountain (stop 1-stations 1 through 5).

Granite. A granite stock approximately 1.7 km in diameter intrudes the southeast portion of the main syenite stock at Ascutney Mountain. The granite is medium- to coarse-grained and subporphyritic. Its major constituents are microperthite, orthoclase, albite, quartz and biotite. Hornblende is rare. Apatite, zircon, sphene, magnetite and ilmenite are the accessory minerals. Chapman and Chapman (1940) mapped an elongate stock of fine-grained hornblende granite, approximately 0.4 km long, on the north slope of Ascutney Mountain. It resembles the surrounding syenite but has approximately 25 per cent quartz. It was not mapped by Daly (1903) probably because he regarded it as a quartz-rich phase of the syenite (stop 4).

Dike Rocks. Daly (1903) mapped four types of dikes at Ascutney Mountain: windsorite, paisanite, muscovite aplite, and lamprophyres. Balk and Krieger (1936) recognized devitrified felsites containing spherulites.

Daly defined the rock type "windsorite," so named for Windsor, Vermont, to describe a leucocratic variety of quartz monzonite with a small amount of biotite. The rock is characterized by high alkali content (K>Na), low Ca (contained only in plagioclase), Fe and Mg. Two windsorite dikes, which are light-colored and 0.3 m to 1 m in width, cut the eastern portion of the gabbro-diorite stock. They consist of plagioclase ($An_{10-50}$), orthoclase, microperthite and biotite. Some of the plagioclase grains are rimmed by alkali felspar. Quartz occurs interstially and augite and hornblende are present though rare.

Paisanite dikes cut the gabbro-diorite and syenite ring dike of Little Ascutney, and the main syenite stock. On Little Ascutney, the paisanite dike has a maximum width of approximately 40 meters. The paisanite dike cutting Mt. Ascutney has two forks each of approximately 20 meters width. Daly noted that these dikes strikingly resemble fine-grained phases of the main syenite stock. They are fine-grained and slightly porphyritic consisting of microperthite, sodic-orthoclase, quartz and alkali-rich hornblende. Microperthite and orthoclase occur as phenocrysts. Hornblende in the groundmass is poikilitic and has resorbed rims. Sphene, ilmenite, zircon, and apatite are accessory

minerals. The paisanite dikes also resemble the syenite in the presence of abundant basic segregations and "kernels" of hornblende and biotite (stop 1-station 6).

Muscovite aplite cuts the main syenite stock. It consists of quartz, orthoclase, albite, some microperthite and muscovite and has a sugary texture typical of aplites. Miarolitic cavities are common and contain terminated quartz crystals and muscovite plates.

Lamprophyres cut the country rock, gabbro-diorite, and syenites but not the granite stock. Two types of lamprophyres occur. Camptonite dikes intrude the gabbro-diorite and syenite of Little Ascutney and the main syenite stock. They are dark-colored and fine-grained. Phenocrysts of feldspar and hornblende are seen rarely. The rock consists of labradorite and hornblende and is altered to a mixture of chlorite, epidote, calcite and quartz. Euhedral basaltic hornblende phenocrysts are 1-3 mm in length and rare augite is altered to uralitic amphibole and chlorite (Daly, 1903). Diabase dikes like those observed throughout the Connecticut River valley consist of labradorite and interstitital augite. Both the plagioclase and the augite are altered to uralitic amphibole and chlorite (Daly, 1903). Pyrite and titaniferous magnetite are the main accessory minerals (stop 3).

Aphanitic, felsite dikes cut country rock gneisses, gabbro-diorite and syenite of Little Ascutney Mountain (Balk and Krieger, 1936). They have the same mineral assemblage as the paisanite dikes and are probably related to them (Balk & Krieger, 1936). They are distinguished from the paisanites by the presence of flow-banding, spherulites and an originally glassy groundmass. The spherulites occur in clusters that are roughly parallel to the flow-banding and are composed of microlites arranged in a radial fashion (Balk & Krieger, 1936). The original glassy groundmass is highly devitrified.

Foland et al. (1985) proposed that the intrusive rocks of the Ascutney igneous complex were derived by fractional crystallization of batches of an alkali basalt parent magma. Their model for crystallization of the mafic and granitic rocks of the complex was based on $^{87}Sr/^{86}Sr$ ratios and $d^{18}O$ values for gabbro-diorites and granites. Their data indicate that the magma that crystallized to form the gabbro-diorites was a mantle-derived basalt considerably contaminated by assimilation of overlying Precambrian gneiss. Their isotopic data for the granite suggest that it was derived from basaltic magma by fractionation with little or no crustal assimilation. Their model does not consider evolution of the syenites. However, field evidence in the form of angular country rock inclusions in the syenite at Crystal Cascade (stop 2) suggests that the syenite magma stoped but did not assimilate crustal material. A crystallization history involving fractionation of early-formed mafic constituents and relative iron and alkali enrichment is indicated by variations in mineral and whole rock chemistry (Schneiderman, 1987).

## BRECCIA XENOLITHS OF LITTLE ASCUTNEY MOUNTAIN

Enclosed in the syenite porphyry ring dike at the top of Little Ascutney Mountain are at least fifteen large, mappable xenoliths of breccia as well as numerous, small, unmappable breccia fragments. The presence of these breccia bodies was first noted by Hichcock (1861) in his report on the geology of Vermont; he regarded them as metamorphosed conglomerates and the remains of more extensive conglomerates which melted to form the "granitic" rocks of Ascutney Mountain. Van Hise (1890) disagreed; thinking the matrix was igneous, he considered the bodies to be flow breccias and the contained rock fragments to be representative of the country rocks through which the magma matrix passed. Daly (1903) found that the matrix consisted of metamorphic mineral and rock fragments. Suggesting that the matrix formed by the comminution of larger rock fragments in the breccia, he considered the breccia bodies to be fault breccias formed in an early fracture zone and later picked up by the intruding syenite magma. Chapman and Chapman (1940) suggested that the breccias were formed by maar-diatreme explosions which shattered the above-lying country rock.

Figure 4 is a small scale map of the top of Little Ascutney Mountain showing the location of the fifteen largest xenoliths of breccia and the stations to be visited during stop 1 of the field trip. The largest xenolith is approximately 50 m x35 m and the smallest one mapped is approximately three meters square. All of the breccia masses have sharp contacts with the enclosing syenite porphyry and appear as irregularly- shaped bodies in it. Thin apophyses of syenite porphyry perforate the breccia xenoliths at the contacts and narrow dikes cut portions of large breccia masses.

Sparse vegetation on the breccia masses leaves them well-exposed as south-facing ledges at the top of Little Ascutney. All of the rock fragments in the breccias are angular or subangular. The rock fragments in the breccias

range in size from one centimeter to two meters in diameter. The breccia xenoliths differ in the sizes of rock fragments they contain; while some of the breccia masses contain rock fragments of greatly disparate size, others have mostly small rock fragments of uniform size. In terms of proportions of rock fragments in the breccias, however, all of the masses are approximately the same; point-counting on the outcrop reveals that rock fragments larger than 2 cm x 2 cm in size account for approximately 43% of the breccia masses.

A great variety of rock types is represented as fragments in the breccia xenoliths: chlorite biotite quartzite, chlorite garnet quartzite, graded- and cross-bedded pelitic metasandstone, garnet biotite schist, garnet cordierite schist ±andalusite ±sillimanite, corundum garnet cordierite schist ±andalusite ±sillimanite, quartz cordierite gneiss ±sillimanite ±corundum ±spinel, trachyte, amphibolite, calc-silicate granulite, chalcedony, and vein quartz. The pelitic fragments are most abundant. Detailed petrographic descriptions of these fragments are given by Schneiderman (1987).

The matrix of the breccia xenoliths consists of broken and comminuted pieces of the fragments in the breccia and their constitutent minerals. Most abundant as porphyroclasts are angular quartz fragments, aggregates of biotite, euhedral fragments of garnets, and plagioclase displaying bent or broken twin lamellae. Extremely fine-grained rock flour made of the pulverized fragments constitutes the bulk of the matrix.

Mineral assemblages and textures indicate that most of the rock fragments in the breccia xenoliths have obvious sources in the Paleozoic metasediments exposed east of the Chester dome and west of the Monroe line in the Ascutney contact aureole. However, evidence including 1)primary sedimentary features preserved in the metasandstone fragments, 2)localities of know occurrences of similarly bedded units in the New Hampshire sequence east of the Monroe line, and 3)metamorphic grade of the quartzites, indicates that the sources for the metasandstone and quartzite fragments were Siluro-Devonian units of the New Hampshire sequence, e.g. Littleton formation (Schneiderman, 1987, 1988). Country rock source units will be seen at stops 2,3 and 6 on the field trip. This implies that New Hampshire sequence rocks occurred on top of Vermont sequence rocks during the Cretaceous; they were available for incorporation as fragments in the breccia during the phreatomagmatic explosion that shattered the overlying country rock and formed the breccia. This interpretation is supported by the size distribution of rock fragments in the breccias. Fragments whose source was New Hampshire sequence metasediments would have traveled the shortest distance in the breccia-forming explosion since these units would have been at the top of the stratigraphic section; these fragments are the largest fragments in the breccia (Schneiderman, 1987, 1988). The interpretation is also supported by geobarometric data. A variety of geobarometers applied to the pelitic rock fragments in the breccia indicate the lowest pressures, less than 2kb, and therefore shallowest depth of origin for the pelitic metasandstone fragments (Schneiderman, 1987,1988). Thus, the breccia xenoliths at Little Ascutney Mountain contain crucial geologic evidence for the presence of New Hampshire sequence rocks above Vermont sequence rocks at least as late as the Cretaceous.

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## ROAD LOG

The assembly point is the commuter parking lot on the south side of Vermont route 131 west of interstate 91, exit 8 at Ascutney, Vermont. The assembly time is 8:00.

Mileage

Quadrangle map: Cavendish, Vermont 1:24,000.

0.0 Exit commuter parking lot (park and ride) and turn left heading west on route 131.

0.2 Pass power station on south side of route 131.

0.5 Pass Victory Drive on the north side of route 131.

2.0 Pass Weathersfield Center Road to the south.

3.0 View of Mount Ascutney (summit 1048 m) to the north.

3.1 Pass Gulf Road to the south.

3.2 Pass Cascade Falls Road to the north.

3.4 Pass Gravelin Road to the south.

4.1 View of Little Ascutney Mountain (summit 570 m) to the north.

4.8 Turn right onto Little Ascutney Road (road is unmarked; red farmhouse is located at the intersection of Little Ascutney Road and route 131) and park vehicles along the right side of the road.

**STOP 1**. RING DIKE OF LITTLE ASCUTNEY MOUNTAIN.

We will spend at least one hour exploring intrusive and extrusive features associated with the ring dike. The outcrops we will examine are located near the top of Little Ascutney Mountain. There is no trail up this mountain; the most expeditious route is N20W through the cow pasture behind the red farmhouse at

Figure 3. Generalized geologic map of the Ascutney complex and surrounding country rocks (Chapman & Chapman, 1940; Nielson, 1973; Thompson, unpub. data).

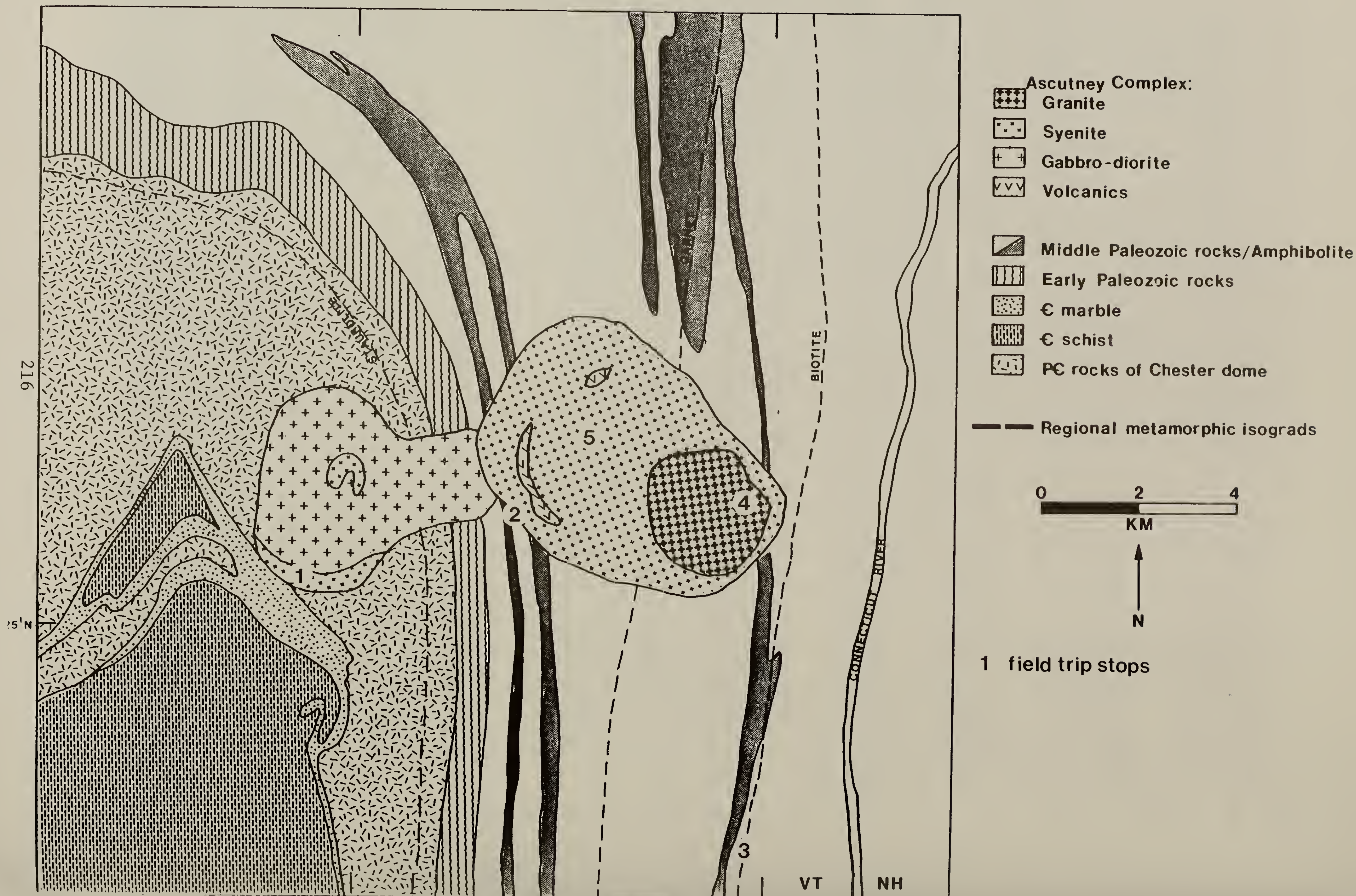
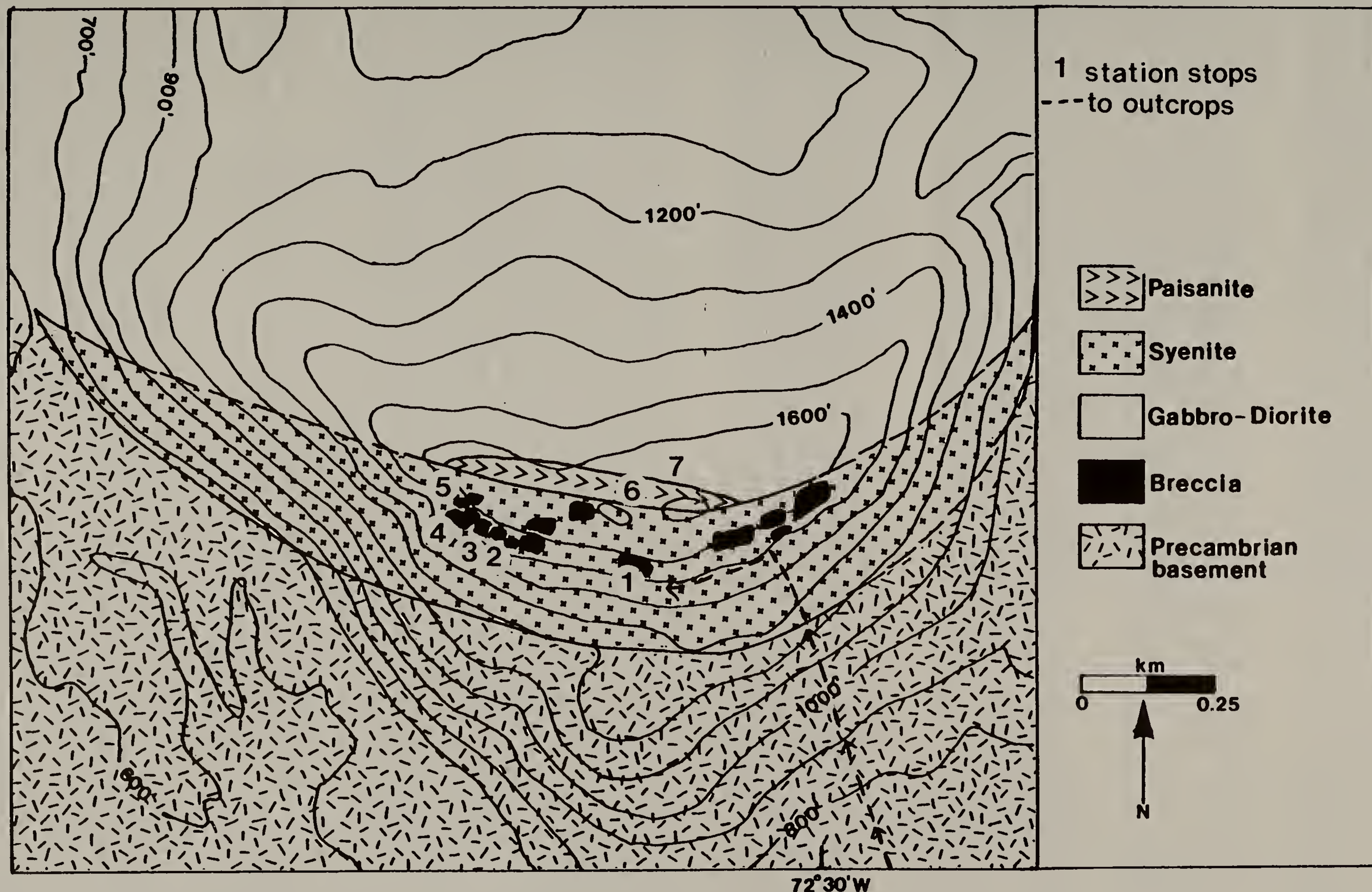


Figure 4. Map showing locations of fifteen largest breccia xenoliths in syenite ring dike, Little Ascutney, Vermont.



the intersection of Little Ascutney Road and route 131. Please request permission from the landowners. Outcrops of Chester dome core gneisses occur in the pasture. Excellent view of Mount Ascutney to the east from the cow pasture.

Walk up the second hill in the pasture at N20W. At the limit of the pasture and the beginning of the forest walk in a direction approximately N60W. Cross a barbed-wire fence at the northern limit of the pasture and continue north through the open woods. As the grade steepens, walk N65W around cliffs to the top of the mountain.

Figure 4 is a sketch map of Little Ascutney Mountain showing station stops. Walk west along the south side of the mountain to station 1. Station 1 is recognizable by the following characteristics: unobstructed view of valley to the south; smooth outcrop surface marked by drill holes at the eastern end; sharp drop at southern end of outcrop. PLEASE, NO HAMMERS AT STATIONS 1 THROUGH 5; the textures that can be observed on these well-exposed, weathered surfaces will be destroyed by hammering.

Station 1. Looking south from this vantage point you can see Hawks Mountain marking the northern portion of the Chester dome (S55W), Pine hill, a north-south trending ridge east of Hawks Mountain (S35W), and on a clear day, Mt. Monadnock (~S10W). The outcrop is a breccia xenolith approximately 35 m x 45 m in size. Three extraordinary rock fragments displaying graded bedding, one as large as 0.7 m x 1.3 m, can be see in the breccia near the eastern edge of the breccia mass. A variety of different rock types (marked by drill holes) can be seen on the southern cliffs of the exposure. Walk N35W approximately 300 meters to station 2.

Station 2. There is an unobstructed view to the south from the outcrop. Near the northern and eastern ends of this outcrop a contact between a mass of breccia and the enclosing syenite is well-exposed. Note the irregular but sharp nature of the contact and the lack of any thermal effects at the edge of the breccia xenolith. The syenite is porphyritic and consists of perthite, hornblende and biotite with minor amounts of quartz, augite, apatite, zircon, sphene and iron oxides. Walk N20W approximately 30 meters to the next clear exposure with unobstructed view to the south.

Station 3. On this well-weathered outcrop surface a diverse array of rock types can be discerned as fragments in the breccia xenolith. Amphibolite is dark green, subangular, and forms depressions in the outcrop surface. Calc-silicate is pale green, angular and commonly is surrounded by a white brecciated rim. Quartzites occur as angular, gray fragments that protrude from the weathered surface. Metasandstone fragments display graded- and cross-bedding. Garnet biotite schist fragments contain 2-6 mm diameter garnet porphyroblasts, and show blackened lineations due to the presence of quartz, cordierite, and corundum. Andalusite or sillimanite, though not macroscopically visible, are present in some of the schist fragments. Gneiss fragments display prounounced banding resulting from alternating layers of massive quartz and cordierite, plagioclase and alkali feldspar. Trachyte fragments are subangular, even-grained and weather brown. It may be difficult to distinguish trachyte from quartzite fragments but the former are softer. At the northeastern end of the outcrop apophyses of syenite can be seen to have intruded the breccia xenolith. Walk N10W approximately 50 meters and down in elevation approximately 15 meters to the outcrops of station 4. (View to the south is not well-exposed.)

Station 4. At the eastern end of this weathered outcrop area is a 0.5 m x 1 m corundum-bearing gneiss fragment (marked by a series of 7 drill holes.) At the western edge is a 0.7 m x 0.7 m metastandstone fragment with well-preserved graded bedding. Each bed is approximately 13 cm thick. Walk N40W approximately 30 meters to station 5.

Station 5. This is the westernmost exposure of breccia enclosed in the syenite. Unlike the breccia xenoliths at the previous stations, this mass consists of many small rock fragments of uniform size. There is a clear view N65W across the rocks of the northern portion of the Chester dome to Okemo Mountain east of the Green Mountain anticlinorium. Walk S80E back towards station 1.

Station 6. Near the middle summit of Little Ascutney north of station 1 are outcrops of a paisanite dike. The rock is fine-grained and consists of microperthite, orthoclase, quartz and alkali-rich hornblende (A-site occupancy=0.70). Such felsite dikes, including some with spherulites and devitrified groundmass (Balk &

Krieger, 1936), intrude the gabbro-diorite of Little Ascutney and syenites of Ascutney Mountain. Continue walking approximately S80E and towards the north slope of the mountain.

Station 7. Outcrops of gabbro-diorite occur along the top of Little Ascutney Mountain and down the north slope. The rock consists of plagioclase, augite, biotite, hornblende, alkali feldspar and quartz; zircon, sphene, apatite, and iron oxides are the accessory minerals.

Return to the vehicles. Continue west on Little Ascutney Road.

5.2 Bear right at fork.

5.8 Turn left and head south on Lottery Lane (unmarked; yellow house at right corner.)

6.4 At the intersection with route 131 turn right and proceed west towards Amsden.

6.7 Turn right on Amsden School Road heading northwest.

7.1 At the intersection with route 106 opposite the Weathersfield landfill turn right and go north.

7.7 At the intersection with Little Ascutney Road turn right headed east (sign for Fuller's Gardens.)

8.0 Pull over on right side of the road for an excellent view of cliffs of stop 1 to the north.

8.3 Turn right and go south on Lottery Lane.

9.2 At the intersection with route 131 turn left and go east towards Ascutney Mountain.

10.2 Pass intersecton with Little Ascutney Road to the north.

Quadrangle map: Claremont, N.H.-Vt. 1:62,500 or Mt. Ascutney, Vt. 1:25,000.

10.8 Pass Ascutney Notch Road to the north,

11.6 Pass Gravelin Road on right.

11.8 Turn left and go north on Cascade Falls Road.

11.9 Turn left and go west on High Meadow Road.

12.3 Park on the right side of the road at trailhead of the Weathersfield trail to Crystal Cascade.

STOP 2. CONTACT AUREOLE AND QUARTZ SYENITE OF ASCUTNEY MOUNTAIN ALONG THE WEATHERSFIELD TRAIL.

The Weathersfield trail is being reconstructed; at the time of this writing, thetrail begins east of the stream, crosses the stream at low elevation (370 m) and continues along the west side of the stream. Follow the Weathersfield trail north up Ascutney Mountain to the waterfall (approximate elevation 500 m) which marks the contact between the intrusion and country rocks.

The trail traverses pelitic schists and calcareous interbeds of the Waits River and Gile Mountain formations. Note from figure 3 that these units are separated by a canoe-shaped loop of the Standing Pond amphibolite. Inside the contact aureole, pelitic units of the Waits River and Gile Mountain formations have a bluish gray color and consist of cordierite, plagioclase, potassium feldspar, quartz, and biotite. Garnet and andalusite occur as porphyroblasts. As the contact is approached, the pelites become increasingly harder and finer-grained hornfelses. Within 25 m of the contact, fibrolitic sillimanite, pleonaste, and corundum are observed. Calcareous units in the contact aureole have a greenish hue and consist of diopside, quartz, wollastonite, grossular garnet, plagioclase, scapolite, sphene, and carbonate. These units are likely sources of the garnet biotite schist, garnet cordierite schist, and calc-silicate fragments in the Little Ascutney breccia.

The quartz syenite at Crystal Cascade consists of perthite, Fe-rich hornblende and annite, with minor amounts of quartz and fayalite. Many xenoliths in the syenite can be seen in the polished outcrop near the crest of the waterfall. Most abundant are subrounded, mafic inclusions that are roughly 10 cm x 15 cm in size and that occur in clusters. They consist of perthite, microcline, orthoclase, plagioclase, hornblende and augite phenocrysts in a fine-grained matrix of plagioclase, hornblende, biotite and quartz. Foland et al. (1985) suggest that these inclusions were stoped from overlying cogenetic volcanic rock as the magma was emplaced. Angular xenoliths of country rock in the syenite can be seen in exposures in the brook approximately 100 m north of the lip of the waterfall. Some of the blocks are as large as 1 m x 1.5 m. It is the presence of such xenoliths in the syenite at Ascutney, particularly at this locality, that led Daly (1903) to formulate his theory of magmatic stoping as a mechanism for the emplacement of intrusive bodies.

Approximately 250 m west of Crystal Cascade at an elevation of 520 m is the contact between syenite and the western band of the Standing Pond amphibolite unit. Within the contact aureole the unit is comprised of hornblende, diopside, biotite, scapolite, plagioclase and garnet. The amphibolite displays a deformed garbenschiefer texture, sprays of 5 cm long hornblende blades radiate from euhedral 2.5 cm diameter garnets, approximately 100 m south of the contact with the syenite. The unit is a possible source of the amphibolite rock fragments in the Little Ascutney breccia.

Return to vehicles. Go east on High Meadow Road.

12.7 Turn right at the intersection with Cascade Falls Road and go south.

12.8 At the intersection with route 131 turn left and go east.

13.9 Wheeler Camp Road to the north.

15.7 Power station to the south.

15.9 Park in the commuter parking lot (park and ride) on the south side of route 131.

STOP 3. SCHISTS OF THE GILE MOUNTAIN FORMATION.

Cross route 131, carefully, to examine the outcrop on the north side of the road. These rocks are typical of Waits River and Gile Mountain formation metasediments outside of the Ascutney contact aureole. The outcrop consists of mica schists, quartzites, and calcareous schists. The mica schists are gray, medium-grained, and consist of quartz, feldspar, muscovite and biotite. They are interlayered with micaceous and feldspathic quartzites. Small, rusty-weathering pits in the quartzite indicate the presence of a carbonate. The calcareous schists weather brown, are fine-grained, and contain thin beds of punky, brown-weathering impure marble. On the east side of the outcrop, along the exit from the interstate, a 2 m wide mafic dike cuts the metasediments. The dike is vesicular and has 15 cm wide chilled margins.

Return to vehicles. Turn right onto route 131. Pass underneath interstate highway 91.

16.5 At the intersection of route 131 with route 5 turn left and proceed north towards Windsor.

17.1 Fork left onto route 44A towards Brownsville.

18.6 Cross over interstate 91.

18.8 Turn left at entrance into Mt. Ascutney State Park, and proceed west up the motor road (use low gear to avoid engine overheating).

19.6 Pull over on right side of road along outcrops with drillholes.

STOP 4. GRANITE OF ASCUTNEY MOUNTAIN.

The granite of Ascutney Mountain, which forms a stock intruding the syenite, is comparable to the Conway granite so abundant in the White-Mountain Plutonic-Volcanic series rocks of New Hampshire.

Its main constituents are microperthite, orthoclase, albite, quartz and biotite. The accessory minerals are apatite, zircon, sphene and iron oxides. Hornblende is rarely observed.

Continue driving up the mountain.

20.4 More outcrops of granite on right hand side of road.

20.7 Outcrops of syenite on right hand side of road.

22.7 Continue driving up the mountain to the parking area near the summit and park.

STOP 5. SUMMIT OF ASCUTNEY MOUNTAIN.

From the parking arealooking west one can see very clearly the crescentic shape of the ring dike of Little Ascutney Mountain. Also, the more rapid weathering of the gabbro-diorite of Little Ascutney is evident from the bowl shape of the north side of the mountain. Optional: Hike up the trail from the parking area to the summit (880 m) of Mt. Ascutney. Outcrops of syenite occur along the trail and at the top of the mountain.

Return to vehicles and proceed down motor road (use low gear.)

23.6 Turn left into scenic area. View of New Hampshire sequence rocks of the Bronson Hill anticlinorium to the east.

23.7 Exit scenic area and turn left onto motor road. Proceed down road to exit of the state park.

26.7 Turn right onto route 44A towards Ascutney.

27.9 Join route 5 and proceed south.

29.1 Turn right onto route 131 going west.

29.6 Turn left onto entrance ramp for I-91 south.

46.2 Take Rockingham exit (#6). At bottom of exit ramp go right onto route 5 south towards Bellows Falls, Vermont.

49.6 At fork bear right (stay on route 5).

49.9 Pass School Street on the left.

50.4 At the traffic light turn right onto route 121 heading west towards Grafton.

Quadrange map: Bellows Falls, Vermont 1:62,500.

51.0 Pass Forest Street on the left.

51.2 Pass Oak Hill Terrace on the right.

51.5 Turn left at small circle onto Gage Street (unmarked).

51.6 The road becomes dirt, turns left and heads downhill.

51.9 Park vehicles on left side of the road near the bottom of the hill before the private homes. Walk to the river across the road.

STOP 6. LITTLETON FORMATION ALONG SAXTONS RIVER.

The outcrops at this locality show the typical graded bedding of the Devonian Littleton formation of the New Hampshire sequence. The Littleton formation is the most likely source of the rock fragments with graded bedding in the Little Ascutney breccia xenoliths.

Clots of chlorite and muscovite in the aluminous portions of the graded beds are pseudomorphs after staurolite. Note that since the rocks display reverse graded bedding, tops are to the west. These rocks belong to the lower (inverted) limb of the Skitchewaug nappe of western New Hampshire (see Thompson et al., 1968 and Thompson and Rosenfeld, 1979).

Return to vehicles. Proceed straight ahead on this road, past homes on the left. The road goes uphill and to the left.

52.5 Turn right onto route 121.

53.0 Proceed straight through the traffic light. Do not turn onto route 5.

53.3 Turn right to bridge over the Connecticut River and to route 12 towards Keene, New Hampshire.

End of field trip.

# SNOWBALL GARNETS REVISITED, SOUTHEAST VERMONT[1]

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## ABSTRACT

In the forensic study of tectonometamorphism, it is difficult to imagine a device better designed to record the sequence of that process' features than a garnet in a schist. With its inclusions, composition, thermoelastic effects around inclusions, and low diffusivities for its constituent ions, that mineral preserves a shell-by-shell record not only of mineral assemblage and fabric sequence but also, with caution because of non-equilibrium effects, of pressure-temperature conditions at time of inclusion. In particular the so-called "snowball" fabric records the syncrystalline rotational motions in terms of their sequence, their directions, and their magnitudes.We now find that the garnet shells also record quantitatively the *rates* (and probably the actual *times* ) at which the successive shells accreted in the form of a radial increase in their initial $^{87}Sr/^{86}Sr$ values. We will demonstrate at least three major garnet-related tectonometamorphic epochs consisting of: (1) a pre-Acadian poorly documented epoch, possibly pre-Taconic, near the Cambro-Ordovician boundary around an approximately east-west tectonic axis (referred to present geographic coordinates); (2) Acadian nappe formation; and (3) dome and easterly backflow development. We will also deal with the implications of continuing geologic mapping in the area and those of current isotopic studies.

## TRIP INFORMATION

The trip will depart promptly from the Keene State College Commons, 8:30 a.m., Saturday, October 15, 1988. **Please sign in so that we don't lose track of you**! We should arrive at Stop 1 on Vermont Route 11, 200 feet west of the pass summit, between Londonderry and North Windham about an hour later. **Participants should provide own sack lunches.** Maximum number of participants is 36. Because some of the trip will involve maneuvering on narrow dirt roads with limited parking space, participants should car (van)-pool to the maximum extent possible! **The trip is a "no-hammer" excursion except for the stop at the Townshend Dam, where excellent samples of snowball garnets are available in road outcrop for unlimited collecting. We'd like to enlist the assistance of all participants in preserving the highly visible minor structural features, thereby enabling future geologists to see them in their field context.**

## INTRODUCTION

The trip's primary purpose is to demonstrate the field component of utilization of rotated garnets in interpreting both history and processes for the tectonometamorphic features of southeast Vermont. While the trip route conforms largely to 1972 NEIGC trip 7 (**Figure 1**, modified after Figure 14-1 of Rosenfeld, 1968), new developments affect the way the outcrops are viewed. It is assumed that attendees will have some familiarity with publications bearing on the trip (Rosenfeld, 1968; Bean 1953; Doll et al., 1961; Rosenfeld, 1970; Hepburn et al., 1984; also Rosenfeld and Eaton, 1985).

[1]Isotopic study reported is supported by NSF Grant EAR 87-07356 and UCLA Grant 1601.

Figure 1, B-6

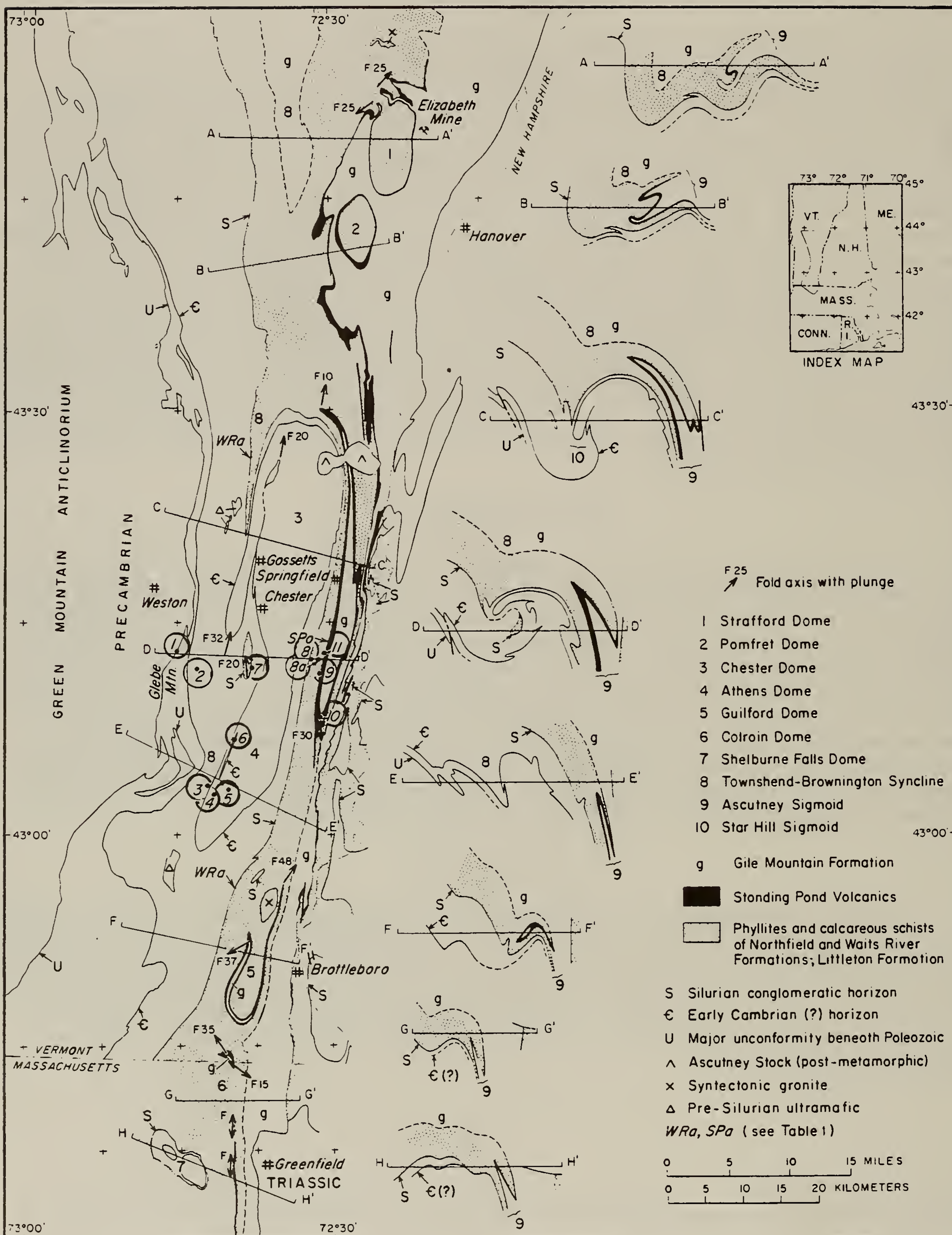



Fig. 1 - Field trip stops italicized.

## USE OF RUBIDIUM-STRONTIUM SYSTEMATICS IN DETERMINATION OF GROWTH-RATE

Among the new developments in particular, requiring a little elaboration, is our developing capability, using Sr-Rb systematics, to measure the growth-rate of large garnets and probably the times at which they grew. Besides the ability to measure isotopic ratios sufficiently accurately, the conceptual model itself appears to be in rather close accord with strict demands of Nature, namely that

1) garnet formed as an accreting, rigid, equant mineral, that crystallochemically excluded all but a trace of Rb and subsequently suffered no significant diffusion of Sr or Rb at the temperatures of concern.

2) garnet grew within a micaceous reservoir matrix that, in turn, sequestered and concentrated a sufficiently measurable amount of the trace element, Rb, within the large alkali positions of the micas, muscovite and biotite.

3) the matrix reservoir near garnet behaved as a closed system with respect to exchange of Rb and Sr with the environment outside the reservoir.

4) *except* for changes due to radioactive decay of $^{87}Rb$, the matrix behaved like a well-stirred isotopically chemostatic bath for the relevant isotopes, Rb and Sr, during garnet growth.

One of the isotopes of Rb, $^{87}Rb$, decays radioactively to $^{87}Sr$. There is also always present a non-radiogenic reference isotope, $^{86}Sr$, which, for the assumed model, stays constant within the system, garnet plus reservoir. Thus the growing garnet, which is chemically unable measurably to distinguish between $^{87}Sr$ and $^{86}Sr$ (unlike the case for lighter isotopes), acquires at its growth surface a trace amount of Sr having $^{87}Sr/^{86}Sr$ equal to that of the reservoir. Thus that ratio today is a function largely of distance from the center of nucleation, the measurable $^{87}Rb/^{86}Sr$ in the reservoir at the present, and time. Also $d(^{87}Sr/^{86}Sr)/dt$ = (decay constant)*($^{87}Rb/^{86}Sr$) $\approx$ a constant for a given specimen because of the very slow decay of $^{87}Rb$. Thus, for all practical purposes, $^{87}Sr/^{86}Sr$ evolves as a straight-line function of time during garnet growth. Because two-point isochrons for garnet and matrix so far seem to give pretty good Acadian ages around 390 million years for the one outcrop so far studied in considerable detail, we may have a fairly good tool for dating metamorphic stages involving garnet growth during the prograde phase of metamorphism. We intend to test this possibility thoroughly.

## GENERAL GEOLOGY

Stops will be in the old 15 minute Saxtons River Quadrangle (now available at 1:25,000 as the 7.5'x15' Saxtons River and Townshend Quadrangles) and are indicated on the summary geologic map and structure sections for southeast Vermont, **Figure 1**. **Figure 2**, a geologic map generated as an intermediate step in preparation of his part of G.S.A. Transect E-1 by J.B. Thompson, Jr. (by permission), provides a closer and more nearly up-to-date view of the geology. **Table 1** is the stratigraphic column, adapted and modified from Figure 2-1 of Hepburn et al. (1984). A principal contrast between **Table 1** and Figure 2-1 is the correlation of the Northfield Formation with the Gile Mountain Formation based on the finding by Rosenfeld, about 1,000 feet west of the Saxtons River - Westminster West Road west of Hartley Hill, of a thin discontinuous zone of Standing Pond Volcanic "look-alike" that he tentatively correlates with that unit. Also carried over in Figure 1 is the correlation of the Northfield Formation with the Littleton. That correlation no longer holds. Facing evidence along the contact between the Put-

Sgm
Ɛo Ɛph
Oum
pƐm
pƐ-Ɛh
Ɛph
OSsa
Omm
Obv
pƐbh
Swr
Ssp
Sgm
10
9a
6
1
2
7
8
8a
9
11
Ɛph Ɛo
pƐm
Srm
Dl
Spv
3
4
5
pƐ-Ɛltm
Ɛph
Ɛo
Ɛs
Sn
Db
Oum
Omm
OSsa
Spv
Swr
Dl
5 miles
After J.B. Thompson, Jr.
43°
42° 45'
72° 45'
72° 30'


Figure 2, B-6

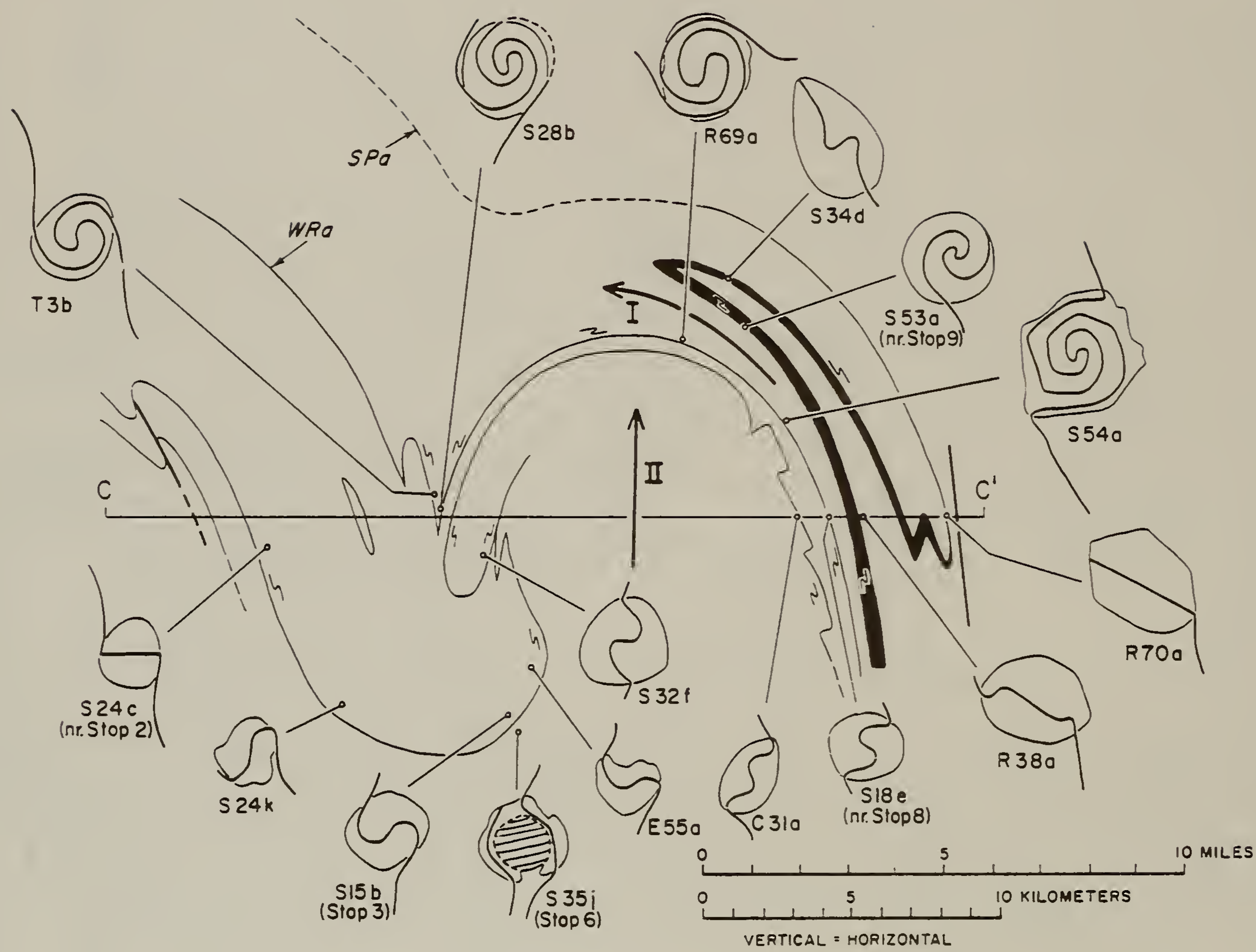
ney Volcanics, previously correlated in Figure 1 with the Standing Pond Volcanics, indicates that the Littleton is younger than the Putney Volcanics. The Putney Volcanics are probably much younger than the Standing Pond Volcanics. Also in the last few years, an important new unit has been broken out as the "Schist-Amphibolite Unit" (Hepburn et al., 1984). Previously that unit had variously been mapped in the area as "Cram Hill Schist," also "Northfield" (cf. the isolated synclinal mass having the dotted pattern on cross section D-D' of Figure 1). The "Schist-Amphibolite Unit" has a conglomeratic quartzite at its base and an immediately overlying sequence not unlike that found in the younger Russell Mountain Formation so well exposed along the Windmill Mountain - Putney Mountain ridge. The isoclinal sigmoid fold in the "Schist-Amphibolite Unit" about 6 miles southeast of Chester does not affect the west contact of the Northfield-Waits River further to the southeast. This is the horizon along which the Russell Mountain Formation with its (basal?) conglomerate is found just to the south. It is tempting to speculate that an important unconformity may lie at the base of that conglomerate. **Figure 3**, with traced patterns from rotated garnets projected into appropriate structural positions on section **C — C'** of **Figure 1**, gives some idea of the way the rotated garnets associate with the kinematics of tectonism in the area. **Figure 4** (= Figure 3-22 of Hepburn et al., 1984; cf. Fig. 14-11 in Rosenfeld, 1968) is a pair of interpretative, sequential, semi-schematic cross sections for cross section **C — C'** to which reference will be made below in referring to garnet rotations, their history, and their tectonic significance. **Figure 4** incorporates the structural implication of the above stratigraphic correlation of Northfield and Gile Mountain in **Table 1**. It should be emphasized that that correlation is not yet "cast in concrete!" (cf. Fisher and Karabinos, 1980, for a different viewpoint).

## SUMMARY OF RELATIONSHIP OF ROTATED GARNETS TO THE GEOLOGY

In a schist of Cambrian (?) age, large rotated garnets have two distinct zones of growth and simultaneous rotation *about axes almost perpendicular to one another, separated by an "angular unconformity"* (see also Karabinos, 1984a,b). The restored early tectonic axis would have been very nearly east-west relative to present geographic coordinates. Combination of that information with the observed absence of such unconformities in much larger snowball garnets of stratigraphically overlying units now thought to be Silurian in age indicated to Rosenfeld (in Zen et al.,1968, p. 196) that the older rocks had already undergone tectonometamorphism, probably in the Ordovician Taconic Orogeny, before the younger Silurian rocks were deposited and subsequently subjected to a new tectonometamorphism during the Devonian Acadian Orogeny. Recent measurements by Christensen of core-matrix two-point Rb-Sr isochrons now indicate that the snowball cores may have crystallized as early as the Cambro-Ordovician boundary in a tectonometamorphic event that may have preceded the Taconic Orogeny. Although less precise, the $^{39}Ar/^{40}Ar$ laser mass spectrometry of Irwin et al. (1987) on a garnet from the same specimen reinforces the same interpretation. In particular both sets of isotopic measurements reinforce the interpretation that there were **two** tectonometamorphic events recorded by the unconformity garnets. Combined with the above discordance of rotational axes, these data do not accord with the conjecture of A.B. Thompson et al. (1977, p. 1163-1164) that the unconformities could have resulted from "one prograde process" in the Acadian.

Examination of the relict included surfaces in garnets of the younger Silurian rocks showed that the garnets commonly had two relict axes of rotation *not separated by growth unconformities* (Rosenfeld, 1970). The inner, or earlier, axes, characteristic of Acadian diastrophic event I, are rotated out of the schistosity in Acadian diastrophic event II by motion about outer, or later, axes within schistosity that are commonly at high angles to the earlier axes. Further, the earlier axes, after graphical correction for rotation about the later axes, are coaxial and cogenetic with large-scale, west-displaced nappes observable in the Silurian units that formed during event I before their arching and folding during development of both the Green Mountain Anticlinorium and structurally lower elongate mantled gneiss domes to the east during event II.

Figure 3, B-6



S28b
R69a
SPa
S34d
WRa
S53a
(nr. Stop 9)
T3b
I
S54a
II
C
C'
R70a
S32f
S24c
(nr. Stop 2)
R38a
S24k
E55a
C31a
S18e
(nr. Stop 8)
S15b
(Stop 3)
S35j
(Stop 6)
0
5
10 MILES
0
5
10 KILOMETERS
VERTICAL = HORIZONTAL

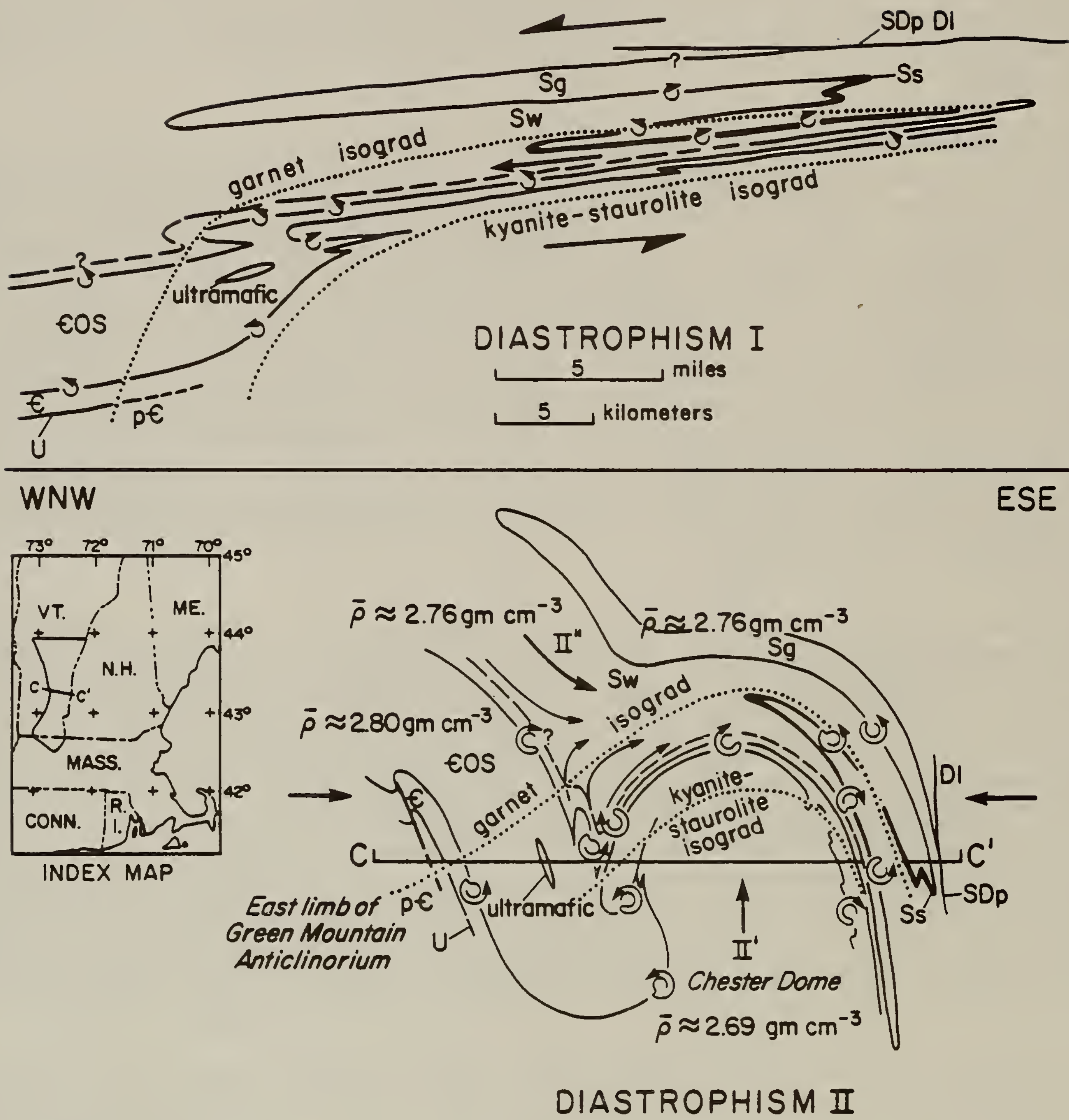
Viewed in a northerly direction (**the convention used unless otherwise stated**), snowball garnets in the lower Waits River Formation on the Chester and Athens Domes rotated in a counter-clockwise direction during event I. Those along the west side of the Standing Pond Volcanics on the west limb of the Ascutney Sigmoid rotated clockwise. Late in event II the garnets and other rotated minerals continued to record the rolling back of the nappes to make backfolds, observed in both Vermont and eastern Connecticut (Hepburn et al., 1984, p. 93-101; Rosenfeld and Eaton, 1985).

The garnet rotations of event I are generally much greater (up to 10 radians) than those of event II (up to 2 radians). The Rb-Sr data indicate that some snowball garnets showing up to 4 radians of rotation that grew during Event I developed about 390 million years ago and took about 10 million years to grow (Christensen et al., 1988). The tectonic history of the two events as interpreted largely from the rotated garnets is shown in **Figure 4**. In **Figure 4** the rotation symbols show the sequence and senses of rotation as resolved into the plane of the section across the Chester Mantled Gneiss Dome and the east limb of the Green Mountain Anticlinorium. It is noteworthy that nappes formed during event I behave as flexural-slip folds in that garnet rotations reverse across their axial surfaces. The pattern of garnet rotations during event II is more complicated. As was the case with event I, rotations reverse across the axial surfaces of the tightly folded synforms near those surfaces. This may be a consequence of flexural-slip folding due to east-west horizontal compression. At deeper levels within the domes, rotations appear to reflect upthrusting of the gneissic cores, possibly due to buoyant forces resulting from the relatively low densities of the gneisses.

With regard to the tectonic significance of stratigraphic facing across the Standing Pond Volcanics, Rosenfeld (1972) had this to say about its significance: "Interpretation of the proximate mechanism of diastrophism for the early and major diastrophic event depends primarily upon knowledge of the as yet unknown age relationship of the units bounding the Waits River Formation on the east. If these units should prove older than the Waits River Formation, the indicated westward transport of material may be ascribed to flexure-slip folding of the westward-opening lower half of a giant lower half of a giant, initially recumbent, sigmoid fold whose upper half is nowhere exposed in eastern Vermont [the interpretation favored here]. If the same units should prove younger than the Waits River Formation, the transport may be ascribed to westward intrastratal extrusion of the relatively dense Waits River Formation, possibly down a gently inclined slope tilted toward the west. It is thus of great importance to resolve this ambiguity by development of procedures for resolving the above stratigraphic uncertainty." This problem, affecting tectonic interpretation of event I, is still a troubling one beset by ambiguous evidence.

At shallower levels within the highly plastic Silurian phyllites and calcareous schists, rotations of garnets appear to be due in part to an *easterly* flow off of the Green Mountain anticlinorium and over the Chester and Athens mantled gneiss domes, shown by streamlines in **Figure 4**. This is in the opposite direction of the earlier *westerly* transport of the nappes that were bowed up by those domes. This signifies that tectonism at a scale even larger than that of the geologically mapped large structures was involved. At that time before the plate tectonic hypothesis (1959), the picture was not very clear as to just what that larger scale process was, although Rosenfeld recalls thinking that it might well have been related to an asymmetric "Benioff-type" gravity pattern then recently observed in the area and associated with thrust faults by R.J. Bean (1953). Bean's interpretation was consistent with the clearly higher structural level of the Green Mountain Anticlinorium relative to the Chester, Athens, and Guilford Domes to the east. At the time of writing of the original study (Rosenfeld, 1968), the possible plate tectonic implications of *retrocharriage* had not yet sunk in. This large-scale flow, extending across the major structures of event II, is like that which must have caused the well-known back-folds in the central Alps, described long ago by Argand. It caused some backfolding in southeast Vermont

## Figure 4, B-6




### EXPLANATION

- DI gray schist
- SDp greenschists (volcanic)
- Sw calcareous schists and phyllites
- Ss greenschists and amphibolites (volcanic)
- Sg quartzo-feldspathic schist, phyllite, calcareous schist
- €OS schists, gneisses, amphibolites
- € augen gneiss, albitic and paragonitic schist, dolomite
- U major unconformity
- p€ polymetamorphic gneisses, schists, amphibolites, marbles

(Hepburn et al., 1984, p. 93-101). The backflow or *retrocharriage* would seem to have resulted from eastward tilting of the lithospheric substrate underlying the anticlinorium and the dome, perhaps a consequence of major and deep westward overthrusting affecting the lithosphere underneath and to the west during event II (cf. Ando et al., ☞∞, 1984). Acadian *retrocharriage* also displays itself well in eastern Connecticut (Rosenfeld and Eaton, 1985). As of the time of writing we have not obtained Rb-Sr evidence as to the timing of event II.

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## TABLE 1, TRIP B-6

### Lithologic Units

Stratigraphic units:

| | |
|---|---|
| **Dl** | Littleton Formation. Monotonous grey slate and schist with thin conglomerate at its contact with the Putney Volcanics. |
| | ——Major unconformity?—— |
| **Spv:** | Putney Volcanics. Buff to light brownish-gray feldspathic phyllite; thin beds of feldspathic granofels; light greenish-gray phyllite; minor gray slate. |
| **Swr:** | Dark gray mica schist and calcareous mica schist with abundant interbeds of punky-brown-weathering, impure marble; thin interbeds of impure quartzite. |
| **Ssp:** | Standing Pond Volcanics. Dark gray to black, medium-grained amphibolite and epidote amphibolite predominant; very coarse grained garnet-hornblende fasciculite |

schist; rare, impure quartzite, coticule and schist (western bands). Gray to greenish, massive plagioclase-biotite-quartz and plagioclase-biotite-hornblende-quartz granofels (eastern band).

| | |
|---|---|
| **Sgm, Sn:** | Gile Mountain and Northfield Formations. Interbedded light gray, impure quartzites and mica schist. Gray, fine-grained phyllite and slate with interbedded, thin, micaceous quartzite and subordinate, punky - brown-weathering impure marble (eastern band). Some interbedded black phyllite. |
| **Srm:** | Russell Mountain Formation. Light gray to white quartz pebble conglomerate and quartzite at and near base; coarse-grained hornblende fasciculite schist; medium-grained amphibolite; silvery schist with tiny euhedral garnets and containing abundant nodules composed of manganiferous garnet, quartz, and apatite |
| | ——Major unconformity?—— |
| **OSsa:** | Schist-Amphibolite Unit. Basal conglomeratic quartzite; rusty-weathering black carbonaceous phyllite and schist; minor gray schist, black amphibolite, black platy quartzite, thin-laminar magnetite-cummingtonite coticule beds common, blasto-porphyritic amphibolite common.There is a distinctive sequence at the base of this unit northwest of Grafton: conglomeratic quartzite; overlain in turn by amphibolite; silvery schist with tiny euhedral garnets and containing abundant nodules composed of manganiferous garnet, quartz, and apatite; thin, buff-white micaceous calcite marble; garnetiferous graphitic schist. |
| | ——Unconformity?—— |
| **Obv:** | Missisquoi Formation, Barnard Volcanic Member. Dark gray to black, poorly to well-foliated amphibolite and porphyritic amphibolite; gray to light gray quartzo-feldspathic gneiss, schist, granofels and layered gneiss; minor black, rusty-weathering mica schist, especially at top northwest of Grafton. |
| **Omm:** | Missisquoi Formation, Moretown Member. Gray to light gray, interbedded quartzite, quartz-mica schist, and impure quartzite. Distinctive laminated "pinstriped" quartzite locally abundant; 10-20% medium-grained amphibolite and garnet amphibolite. |
| **Ꞓs** | Stowe Formation. Garnetiferous chlorite-sercite schist with abundant quartz lenses; subordinate greenish to black amphibolites. Separately mappable only on the east limb of the Green Mountain Anticlinorium |
| **Ꞓo:** | Ottauquechee Formation. Dark gray, rusty-weathering pyrrhotitic mica schist, commonly graphitic, notable for large garnet porphyroblasts; epidote amphibolite; gray to black quartzite interbeds; highly paragonitic near base. |
| **Ꞓph:** | Pinney Hollow Formation. Light green quartz-muscovite-chlorite schist with quartz lenses, commonly paragonitic with thin interbedded laminar epidote amphibolites, especially abundant near top of unit, where they are sometime separately mapped as "Chester Amphibolite." In some places has garnets up to an inch in diameter containing growth "unconformities." Schist occasionally has thin pale pink coticule lenses. |
| **pꞒ-Ꞓh:** | Hoosac Formation. Mica schist with conspicuous albite porphyroblasts and flaggy well-banded gneisses; epidote amphibolite, some with large epidote nodules (where separately mapped is called Turkey Mountain Amphibolite = **pꞒ-Ꞓtm**); silvery paragonitic garnet-muscovite schist with staurolite porphyroblasts and containing garnets up to 1 inch in diameter in some places. |

| | |
|---|---|
| **p€t** | Tyson Formation. Discontinuous polymict conglomerate gneiss at base in some places associated with overlying dolomitic marble; feldspathic muscovite schist. |
| | ——————Major unconformity—————— |
| **p€bh:** | Bull Hill Gneiss. Microcline augen gneiss. |
| | ——————Major unconformity?—————— |
| **p€m:** | Mount Holly Gneiss. Heterogeneous gneisses, commonly quartz-biotite-plagioclase gneiss; some coarse marbles associated with calc-silicates, commonly rusty weathering due to pyrrhotite and containing coarse crystals of graphite. |

Non-stratified rocks:

**Db** Black Mountain Granite. Biotite-muscovite microcline syntectonic granite.

**Oum** Ultramafics occasionally preserving their protolithic character in the form of olivine and pyroxene, but more often altered to serpentine containing abundant carbonate and magnetite in the lower metamorphic grades and soapstone containing talc, carbonates, and various strange biopyroboles in the higher grades.

## ITINERARY

Road log begins on Route 11, just west of the summit of the pass over the east range of the Green Mountains, 0.5 miles southwest of North Windham, Vermont, about 200 feet west of the west boundary of the Saxtons River Quadrangle.

Mileage

0.0 STOP 1. Angular unconformity between a sliver of the overlying prograde metamorphosed basal conglomerate of the Tyson Formation and the underlying retrograde metamorphic rocks and pegmatites of the Precambrian Mount Holly complex. This unconformity is significant for this trip because it demonstrates the direction of stratigraphic "tops." Proceed easterly on Route 11 through the Hoosac Formation. This unconformity was originally noted and mapped by T. Nelson Dale near the beginning of this century (unpublished U.S.G.S. manuscript map). The polymict basal conglomerate is much more convincing at Dry Hill to the north west of Lake Amherst and to the south on the west side of Glebe (Magic) Mountain.

0.6 North Windham. Turn right onto Rte. 121.

0.8 Northernmost exposures of Turkey Mountain Member (amphibolite) of Hoosac Formation outcrop as a sliver only a few feet thick in draw to west. Continue through schists of Pinney Hollow Formation.

2.2 Near crest are exposures of Chester Amphibolite Member of Pinney Hollow Formation. Strong down-dip lineation of pale green amphiboles. Nearby some have been observed by Laird and Albee (1981) to have actinolite cores. Continue through Ottauquechee and Stowe Formations.

2.6 STOP 2 at Lawrence Four Corners. First rotated garnet locality in outcrop at northwest corner of intersection. Garnets in schist of Stowe Formation show small counterclockwise

rotation after growth about nearly horizontal axes when viewed in a northerly direction (the direction of view used subsequently unless otherwise stated). Note effect of garnet rotation on bending of adjacent schistosity. Proceed southerly from Rte. 121 on Windham Road past outcrops of Stowe Formation and rusty schists of the Ottauquechee Formation.

4.0 Windham Center. From here almost to South Windham, the road lies within the banded rusty-weathering graphitic schists of the Ottauquechee Formation. Rise in metamorphic grade is most evident in the field in the transition from pale green amphibolites to dark green to black amphibolites. At the stone bridge at the bottom of the hill immediately south of Windham Center we cross the oligoclase isograd (originally mapped by Rosenfeld, 1954), northwest of which plagioclase more calcic than nearly pure albite is not found, regardless of bulk composition of the rock. Coexisting oligoclase and albite, first observed here using optical immersion methods, can be found in amphibolites just upstream from the bridge. The oligoclase isograd is related to a miscibility gap within the plagioclase feldspar series.

You may be interested in an historical note about the above stone bridge that gives some insight into Vermont, Vermonters, and how times have changed. One day back in the late '40s, an old-timer by the name of Reilly happened to look down from the bridge upon Thompson and Rosenfeld deeply engrossed in the brook outcrop. With an "I gotcha" smile not unknown in these parts, he inquired of those "city-slickers" if they reckoned how much it had cost to build the bridge. Thompson, perhaps harking back to his childhood days among similar folk "down-east" in Maine, paused and then threw out the figure, $50.00. This wiped the smile off Mr. Reilly's face. Somewhat subdued — crest-fallen might be a more apt description — he averred as how he had built that bridge, and that's exactly what it cost! Hence this bridge will ever be known as Reilly's "Fifty-Dollar Bridge!"

7.6 South Windham. Chester amphibolite. In this unit on the ridge just to the northwest, Laird, Lanphere, and Albee (1984, p.387) have obtained an $^{40}Ar/^{39}Ar$ age on amphibole of 376±5.0 million years.

8.0 Jamaica-Townshend town line. Enter the typical green garnet-magnetite-chlorite-sericite schist comprising the main part of the Pinney Hollow Formation and through which the road passes for the next 2.0 miles.

10.0 Turkey Mountain Member appears on ridge to west. From here to West Townshend we pass down-section from the Pinney Hollow Formation into the characteristic albite schists of the Hoosac Formation. To the west of here large isoclinal folds are involved in major thrust faults toward the west (Karabinos, 1984; also Rosenfeld, 1954, geologic map) that probably connect with the fault offsetting the unconformity at the base of the Tyson Formation near STOP 1 that was inferred by Rosenfeld (1954; cf. Figure 2). Not showing on Figure 2 is the extension of the Turkey Mountain Member, traced by Rosenfeld, along Glebe (Magic) Mountain Ridge almost to North Windham; that member is well exposed at the top of the main ski lift at Magic Mountain. A short apparent gap in this unit near Cobb Brook Falls is probably explained by the above-mentioned fault. This fault and others like it may be related to the *retrocharriage* discussed in the text.

11.6 Roadcuts on west side of highway show eastward dipping beds of "pinstripe" in Moretown with a prominent boudinage fracture having horizontal orientation. Continue in highly contorted schists and amphibolites of the Moretown across the axis of the Townshend-Brownington syncline onto the west limb of the Athens (pronounced Aythens) dome.

12.8 Thin amphibolites in smooth coutcrops of Moretown on the left exhibit boudinage.

13.1 Park cars in parking area on right at Townshend Flood Control Dam. STOP 3 is in the roadcut on the northeast side of the highway opposite the dam. The cut is an almost complete exposure of the Ottauquechee Formation, the best and most nearly complete exposure of that unit on either the Chester or Athens Dome. Snowball garnets show counterclockwise rotation on the west limb of the Athens dome. Note the relative consistency of the shear sense indicated by the rotated garnets in contrast to that of the drag folds. The origin of this contrast has been discussed elsewhere (Rosenfeld, 1970, p. 92). Also note the large boudinage fractures in amphibolites here.Garnets observed here are believed to have grown and rotated before development of the Athens dome during Acadian Event I. The relict "oligoclase isograd" may be observed in the form of coexistent albite, oligoclase, and clinozoisite encapsulated in garnets at this locality (Rosenfeld, 1970, p. 90-91), even though the staurolite isograd is only a short distance east. This would seem to be good evidence of growth of the garnets over a considerable range of temperatures. The lowest part of the Ottauquechee Formation, exposed just to the north of the highway about a couple of hundred feet southwest of the dam, is a rusty-weathering paragonite schist having large garnets up to 3cm in diameter. Some of these are the first to have had their growth duration measured. They took about 8 million years to grow as determined from the decay, $^{87}Rb \rightarrow {}^{87}Sr$. Precision is about 2.5 million years. One of these, having the characteristic snowball pattern with rotation of almost 4 radians, thus gives some idea of strain-rate during prograde metamorphism, here about $3 \times 10^{-14} sec^{-1}$. Proceed southeasterly on Rte. 30 through Pinney Hollow Formation amphibolites and schists to the Hoosac Formation.

13.5 Scott covered Bridge on right. Amphibolite in what is believed to be Hoosac Formation on left. If these rocks correlate with the main band of the Hoosac to the west, they are of a distinctly more banded and gneissic facies. Just beyond the bridge on the left are some very nice secondary drag folds on a large fold, incompletely exposed in the outcrop.

13.8 STOP 4. Conglomerate gneiss of Tyson Formation (?) on west in contact with Bull Hill Gneiss, which maps out as a stratigraphic unit on the Chester and Athens Domes. Paul Karabinos and John Aleinikoff have currently been doing U-Pb lead determinations on zircon grains in this unit, getting post-Grenville ages between 900 million and a billion years. Because of the Bull Hill gneiss' granitic composition and its stratigraphic character, it is possible that that unit represents a metamorphosed stack of rhyolitic volcanics. In the southern part of the Athens dome, it has not been possible to delineate accurately the boundary between the Bull Hill gneiss and what are believed to be older but lithologically similar Precambrian granitic augen and flaser gneisses in the core of the dome. Note counterclockwise drag folds in gneiss, believed to be a result of upthrusting of the gneissic core of the dome. Proceed easterly on Rte. 30 through broad zone of granitic gneisses to

15.0 Townshend. Turn left off Rte. 30 onto Rte. 35 and proceed northerly.

15.4 STOP 5. Outcrops lie across the field to the west and consist of magnetite-bearing granite flaser gneiss, believed to have been representative of the relatively low-density "plunger" accounting for the buoyant upward thrust of the Athens dome. Continue north on Rte. 35 through heterogeneous gneisses, some rusty weathering and containing coarse graphite flakes rather like the Washington gneiss described by Emerson in the Berkshires.

16.9 Simpsonville.

18.4 Easy to miss intersection. Bear left off Rte. 35 onto Grafton Road.

18.6 For the next 0.2 miles, passing through a band of calc-silicate rocks, characterized by coarse graphite flakes and pyrrhotite, that strikes northeasterly through the core gneisses of the Athens dome at a large angle to the mantling strata. This discordance provides, perhaps, the best evidence to date that the core gneisses of the Athens dome lie unconformably beneath the mantling strata.

18.8 Continue through banded, contorted, biotite gneisses of the core of the Athens dome.

19.6 Top of grade. Bull Hill Gneiss on dip slopes along east side of South Branch of Saxtons River to north. Valley probably owes its alignment to an easily eroded dolomite (observable at a number of localities on Rte. 35 north of Grafton) that separates albite schist of the Hoosac Formation on the west from the Bull Hill Gneiss.

20.3 Easy to miss turn. Turn sharply left onto single lane, steep dirt road (Acton Hill Road). Proceed through Hoosac Formation.

20.6 Cross brook.

20.9 East contact of garnet-kyanite-straurolite-paragonite schist of Pinney Hollow Formation in core of anticlinal portion of Ober Hill Fold. Pass across Ober Hill Fold.

21.6 Intersection. Let lead car turn around before entering intersection. Then, one by one, each car should turn left, then back up sufficiently far to make room for following cars to do same. Continue back down the Acton Hill Road, following lead car.

21.8 Park your car as far off the road to the right as possible. <u>STOP 6</u>, exhibiting garnets with angular growth unconformities (Rosenfeld, 1968, p. 196), is on the ledges visible to the southwest of the road west of a kyanite-rich paragonite schist. The rock is a garnet-staurolite-paragonite-muscovite schist. Chloritoid and staurolite exist as an armored relict assemblage inside the garnets here. There is no chloritoid outside the garnet. The earlier garnet, showing snowball character, may have grown at about the Cambro-Ordovician boundary possibly during a pre-Taconic tectonometamorphism (see text above). Proceed back toward Townshend-Grafton Road.

22.1. On the left are some remarkably fine counterclockwise drag folds, some of which have transcurrent "slip fractures" of similar shear sense about the same axis. These fractures provide evidence of the "lateness" of these folds.

22.9 Townshend-Grafton Road. Turn left and continue north.

27.7 Grafton, a picturesque village in which some of the finer examples of old Yankee architecture have been restored and preserved by the liberal application of dollars by the Windham Foundation. The local Cheddar cheese from the store we pass in the south part of the village is some of the best made in Vermont. Recently "Funny Farm," starring Chevy Chase, was shot here. You may spot some of the "local characters" that added authenticity to that film. Turn left onto Rte. 121, passing successively through a rather complete section of units from the Hoosac Formation to the rusty-weathering, graphitic schists atop the Barnard Volcanics, formerly correlated with the "Cram Hill Schist" of Jahns, that have not yet received a new name.

29.9 <u>STOP 7</u>. Westward dipping beds of conglomeratic quartzite and interbedded garnet-

muscovite schist at base of Schist-Amphibolite Unit. These beds lie on the east limb of a syncline (Spring Hill Syncline) whose axial surface dips to the west. This syncline was previously believed to be the detached (by megaboudinage) westward-opening, lower part of the Star Hill sigmoid (Figure 1, Section D-D'). Now it is not at all clear how it fits into the multiple tectonism to which it has been subjected and constitutes a puzzle that Jim Thompson, Carl Jacobson of Iowa State, and Rosenfeld have been struggling with. Staurolite and kyanite occur in some of the schists contained within the syncline. The sequence of rotations found in snowball garnets within a schistose parting in the quartzite a mile to the north is: early clockwise, late counterclockwise. Turn around and return to

32.1 Grafton on Rte. 121, continuing through the village across the Saxtons River and turning left onto

32.2 Rte. 35, proceeding northerly along approximately the same stratigraphic horizon that was followed south of Grafton. Bull Hill Gneiss to east.

33.9 Dolomite under albite schist of Hoosac Formation on left. Leaving Athens Dome; entering Chester Dome.

36.4 Enter Grafton Gulf.

36.9 Leave Grafton, Windham County; enter Chester, Windsor County.

37.0 Note pillar of dolomite supporting albite schist on left, dip slope of Bull Hill Gneiss on right.

37.5 Summit of Grafton Gulf.

38.3 Leave Saxtons River Quadrangle; enter old Ludlow Quadrangle (new Chester Quadrangle).

39.5 Chester. Turn right onto Rte. 103.

40.9 Return to Saxtons River Quadrangle.

42.3 Bull Hill Gneiss on east limb of Chester Dome. One of localities dated by Paul Karabinos and John Aleinikoff.

42.5 Enter town of Rockingham, Windham County. Crossing Hoosac Formation.

42.7 Crossing from Pinney Hollow through intermediate units into Moretown Formation.

44.3 Easy to miss intersection. Turn sharp left off Rte. 103 onto dirt road with bridge over railroad tracks.

44.4 Car junkyard on left.

44.5 Covered bridge.

44.9 <u>STOP 8</u>. Ledges in woods north of road. Sieve texture garnets in calcareous schists of lower Waits River Formation showing early counterclockwise rotation (Event I; conspicuous) followed by late clockwise rotation (Event II; observed with difficulty). Continue easterly.

45.7 Optional STOP 8a. Main zone of calcareous schists with subordinate phyllites within Waits River Formation in what used to be one of the best exposures of the Waits River Formation in southern Vermont before recent construction of a power dam. Big sprays of zoisite. Isoclinal folding. Easily observed rotated garnets. Mafic dike with multiple chilled borders and euhedral calcite phenocrysts observable in the chilled borders in thin section. Turn right across bridge and railroad tracks.

46.0 Turn left onto Rte. 103.

46.2 Turn right off Rte. 103 onto Pleasant Valley Road. Passing through heterogeneous rock types of Standing Pond Formation, mostly mafic volcanics.

47.1 Turn right off the Pleasant Valley Road onto single lane dirt road.

47.2 Park cars and proceed northerly across field about 1,500 feet into woods just northwest of northwest corner of field to STOP 9 at contact between garnetiferous phyllite of Waits River Formation on west and coarse garnetiferous schist of the Standing Pond Formation containing sprays of hornblende (fasciculitic schist or "garbenschiefer"). Large garnets show only a single large clockwise rotation associated with Event I, in contrast to those at Stop 8. A photograph of a rotated garnet from this locality appears as figure 14-6 in Rosenfeld (1968, p. 195). Evidence of Event II at this locality appears only as gently northward plunging crinkles. For further discussion of this locality, see Rosenfeld, 1970, p. 89. Return to Pleasant Valley Road by car.

47.3 Turn right onto Pleasant Valley Road.

48.7 Septum of Waits River-like calcareous schist and phyllite in Standing Pond Formation.

48.8 Exposures of banded and massive amphibolites of Standing Pond Formation near eastern contact with Gile Mountain Formation. Clockwise drag folds. Road continues southerly along east side of Standing Pond Formation.

51.0 Intersection with Rte. 121. **Dangerous intersection**!!!! Continue east on Rte. 121.

51.3 Village of Saxtons River. Turn right and go south across bridge over Saxtons River following paved road for about a third of a mile past Kurn Hattin orphanage on left and continuing straight ahead at sharp bend onto Hartley Hill Road (dirt road) continuing up the hill on same.

52.6 About 0.36 miles beyond town line between Rockingham and Westminster just beyond curve to left, turn right onto poorer quality road.

53.0 Turn around and park. STOP 9a is about a hundred feet into the woods east of the road. The horizon of the Standing Pond Formation to be observed is slightly into the unit but otherwise on strike from the outcrop observed at STOP 9. It is not marked on Figure 1 but is just northeast of the "0" in "F30" on Figure 1. This is one of the best places to see in three dimensions the shapes of the included surfaces of the snowball garnets on a weathered steeply dipping dip surface. Here it is easy to see the orientations of rotational axes in the field. Rotational sense is clockwise. Return to

54.7 Saxtons River Village and park just north of bridge. STOP 10. The purpose of this stop is to observe southward plunging minor folds in the Standing Pond Formation along the axis

of the upward closing fold (anticline) of the Ascutney Sigmoid. The axis of this horizon reappears to the south on the Guilford Dome near the syntectonic Black Mountain Granite in Dummerston (fig. 1). Folds with counter-rotating garnets on their limbs appear along the north side of the river, 0.3 miles to the west (Rosenfeld, 1970, p. 85-86). Turn westerly on Rte. 121. For a discussion of the relationship of the rotated garnet history here and at the same structural position on the Guilford Dome as affected by the Black Mountain Granite, see Hepburn et al. (1984).

55.0 Bear right off Rte. 121 onto Pleasant Valley Road.

59.6 Turn left off the Pleasant Valley Road onto Rte. 103.

59.8 Turn right off Rte. 103 toward Brockways Mills, continuing across bridge past Stop 8a and to the right on paved road toward Springfield.

60.9 Park. Proceed westerly across north end of field past small cottage to STOP 11 at contact between garnetiferous phyllite of Waits River Formation and "garbenschiefer" with large garnets. This locality is, perhaps, the best locality for seeing evidence of both Events I and II within a single rotated garnet. A stereoscopic photograph of a rotated garnet you will see at this locality appears as figure 14-3 in Rosenfeld, 1968, p. 192. A discussion of the generation of the central surface of garnets at this locality is found in Rosenfeld, 1970, p. 40. Trip ends here. Return to Keene via Rtes 103, 5 to Bellows Falls, and then 12.

# GRAPHITE VEIN DEPOSITS OF NEW HAMPSHIRE

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## INTRODUCTION

"Black lead (plumbago) is found in large quantities about the grand Monadnock in the township of Jaffrey" (Belknap, 1792, p.195). So wrote Jeremy Belknap in 1792 of the mountain that dominates the eastern horizon of the 80th Annual NEIGC meeting in Keene, New Hampshire. It is not entirely surprising that the early settlers of New England knew of the value of graphite and were able to recognize it in the field. The graphite mines at Seathwaite in Borrowdale of the English Lake District were worked from the 1600's until the 1830's yielding a yearly net profit as high as $92,690 in 1803 (Postlethwaite, 1913). Such large profits would certainly have inspired explorers and settlers to search diligently for the valuable mineral. That the Borrowdale deposits may have served as an "exploration model" for graphite in New Hampshire is confirmed by J. D. Dana who explicitly compared the Bristol, N. H. deposit to the one at Seathwaite (Dana, 1823).

C. T. Jackson reported in 1844 on a number of active workings throughout New Hampshire, "Black lead, or graphite, is a mineral of considerable value to the people of the State, and its sale brings in a constant, though not very great revenue, to those who are engaged in the business. The beds of plumbago are never large, and only employ the farmers during those intervals in agricultural labor, when hands can be spared from other work" (Jackson, 1844, p. 188). Jackson's work remains an invaluable reference for locating ancient workings. Most of the deposits that have been studied are mentioned by him and were pinpointed by reading town history books and interviewing local residents. The history of the deposits is rich in anecdote. Perhaps the most fascinating is the speculation that Henry David Thoreau, whose brief career as a pencil maker coincided with the apex of graphite mining in New Hampshire, may have used graphite from one of the mines to fashion his hand-crafted pencils.

The graphite deposits of New Hampshire no longer have economic value. Their historical significance scarcely deserves so much as a footnote in a textbook. For many years their very existence was forgotten. Today, however, there is increasing interest in the deposits on the part of petrologists and geochemists. Recent research shows that both metasedimentary and metaplutonic rocks of New Hampshire are pervaded by secondary graphite of hydrothermal origin. The prevalence of direct mineralogical evidence of widespread hydrothermal activity throughout the metamorphic belt suggests the important role flowing fluids may play in orogenesis: as a medium of advective heat transfer; as an agent of mass transfer; and as a means of facilitating deformation through hydraulic fracture. The deposits are an example of one of the dilemmas faced by economic geologists. That is, how is it possible to mobilize, transport, and concentrate a mineral so insoluble and refractory as graphite.

The three largest vein deposits known in New Hampshire will be visited on the field trip: (1) Osgood; (2) Franklin Pierce; and (3) Mt. Kearsarge (fig. 1). Below, we give a brief review of published work on general features of the deposits. Following the review, unpublished data on detailed outcrop relations, petrology, and stable isotope geochemistry of the three deposits will be presented

## GENERAL FEATURES

Hydrothermal graphite occurs throughout Silurian and Devonian metasedimentary and metaplutonic rocks of the Bronson Hill anticlinorium and the Merrimack synclinorium (fig. 1). The metamorphic grade of host rocks ranges from kyanite-staurolite at the Keene and Walpole vein deposits to sillimanite for many of the other occurrences. The deposits at Bristol, Antrim, and Sodom Hill occur within granulite facies "hot spots" delineated by concentric sillimanite-alkali feldspar and cordierite-sillimanite-alkali feldspar isograds (Chamberlain and Lyons, 1983).

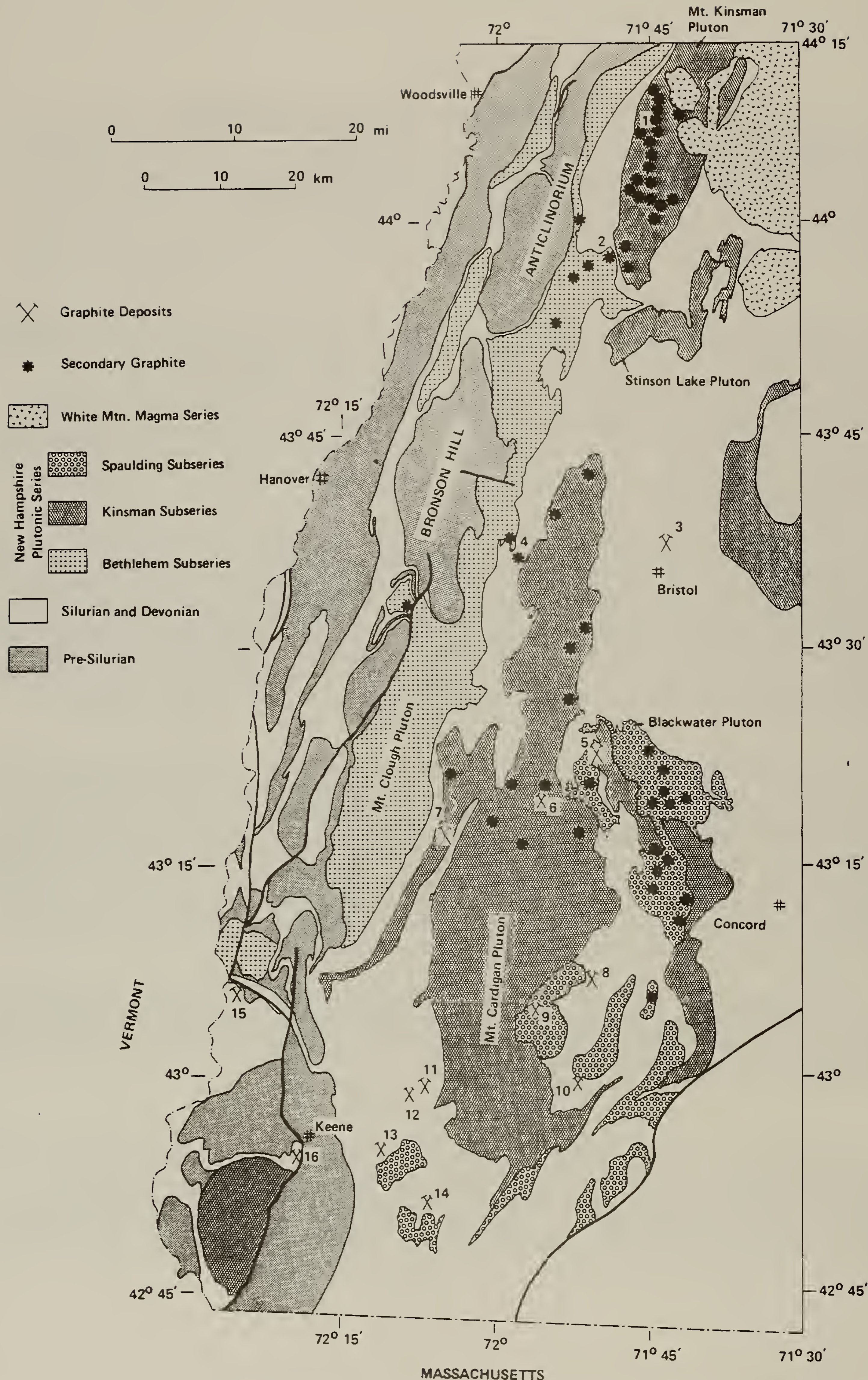


Figure l. - Map of SW New Hampshire showing localities of hydrothermal graphite. Veins are indicated by crossed picks. Field trip will visit localities 11, 7, and 5. Localities are 1 = Eliza Bk. shear zone, 2 = Baker River shear zone, 3 = Bristol mine, 4 = Tewksbury Pond, 5 = Mt. Kearsarge mine, 6 = Sutton mine, 7 = Franklin Pierce mine, 8 = Sodom Hill mine, 9 = Antrim mine, 10 = Crotched Mtn. mine, 11 = Osgood mine, 12 = Nelson mine, 13 = Marlboro mine, 14 = Mt. Monadnock mine, 15 = Walpole mine, 16 = Keene mine.

The geologic age range of hydrothermal graphite lies within the limits of Early Devonian to pre-Jurassic (Rumble, Duke, and Hoering, 1986). The deformed graphite-tourmaline intergrowths of Mt. Monadnock show that some of the veins were emplaced prior to penetrative deformation of the post-Early Devonian Acadian Orogeny. Many veins, however, retain beautiful textures of hydrothermal precipitation and are clearly post-deformation in age. Veins at Franklin Pierce and Bristol probably formed under peak metamorphic conditions as their minerals and wall rock alteration are compatible with prograde metamorphism. At Osgood and Mt. Kearsarge, however, wall rock replacement was accompanied by retrogradation of high grade porphyroblasts. Age dating of zircon overgrowths in the Bristol vein with the SHRIMP ion microprobe gives U-Pb ages of 409 m.y. ($\pm$10 m.y.) (P. K. Zeitler, 1988, personal communication). Metamorphic monazite from wall rocks at Bristol has a conventional U-Pb age of 398 m.y. ($\pm$ 10 m.y.) (B. Barriero, 1988, personal communication). The younger age limit on graphite mineralization is established by two observations: (1) hydrothermal graphite in the Kinsman Quartz Monzonite appears to be bleached out in contact aureoles of Conway Granite in Franconia Notch; and (2) graphite is sheared by a fault whose illite-bearing gouge give a Jurassic K-Ar date (J.B. Lyons, 1983, personal communication)

Hydrothermal graphite occurs in two forms in New Hampshire, as cross-cutting veins visible in the field and as microscopic spherulites and veinlets (Rumble, Duke, and Hoering, 1986). The veins have many features indicating they were formed by precipitation from hydrothermal fluids in fractures. The contacts of graphite veins crosscut bedding, schistosity, and pluton-wall rock contacts. Veins are found at intersections of faults and dikes, along pluton-wall rock contacts and in shear zones. Wall rock alteration zones extend up to 10 cm from vein contacts. The alteration consists of replacement of quartz, micas, and feldspar by graphite or graphite plus tourmaline. Crustiform depositional structures, either concentric in nodules or parallel to vein walls, are found in many deposits. Historical accounts suggest the veins may reach substantial size but never big enough to rival the celebrated veins of Sri Lanka. A large piece of pure "plumbago" measuring 4 ft. x 4 ft x 6 ft. was taken out of the Nelson mine (now flooded) (Elliot, 1941). Jackson (1844, p. 188) reports veins up to 2 ft. thick at Antrim. We have seen veins consisting of quartz and graphite at Bristol that may have extended as much as 20 m in length, before excavation.

Microscopic spherulites, clumps, and veinlets of graphite are the most prevalent forms of hydrothermal graphite in New Hampshire but usually do not exceed 1% (modal) of their host rock. Spherulites are disseminated throughout the pre- to synmetamorphic plutons of the New Hampshire Plutonic series (fig. 1) (Duke and Rumble, 1986). In typical outcrops, spherulites are more abundant in rocks with chloritized botites and serecitized feldspars. Zones of associated retrogradation and graphitization are planar arrays and probably represent healed fractures. Thin sections show that spherulites are oriented in planar arrays, crosscutting feldspar phenocryst. The distribution of secondary graphite given in fig. 1 reflects incomplete mapping in the southern part of the state rather than the true abundance of spherulites.

The Eliza Brook and Baker River shear zones, in the Kinsman and Bethlehem subseries, respectively, have halos of graphite spherulites (fig. 1). The shear zones trend NNE, truncate plutonic rocks, and are filled with a recrystallized mylonite consisting of angular fragments of quartz and serecitized feldspar in a mesh of chlorite, muscovite, and epidote The mylonites are transected by microscopic graphite or graphite-tourmaline veins.

An unexpected and difficult to recognize occurrence has been accidently observed at Tewksbury Pond (fig. 1) along U. S. Rte 4, near Canaan, and near the Montcalm exit along Interstate 89 (courtesy of L. Baumgartner). In both places an alteration zone, from 2 cm to 50 cm thick, enriched in hydrothermal graphite, is found in the wall rock of quartz veins. Curiously enough, the quartz veins themselves have little or no graphite. We suspect that such occurrences may be rather common but have, heretofore, gone unnoticed.

## STABLE ISOTOPE GEOCHEMISTRY

Carbon isotope ratios of graphite from veins, spherulites, and wall rocks overlap in the rang of -28 to -9$^{o}$/oo, $d^{13}C_{PDB}$ (fig.2). Two points should be emphasized in interpreting the data. The first is that both forms of epigenetic graphite, the veins and the spherulites, are similar in $d^{13}C$ to syngenetic graphite from wall rocks. This suggests that hydrothermal graphite shares sources of carbon in common with sedimentary carbon. The second point is that the $d^{13}C$ values of all types of graphite--veins, spherulites, or syngenetic--are intermediate between the two crustal reservoirs of biogenic carbonate minerals (0 $^{o}$/oo) and reduced organic matter (-25 $^{o}$/oo). Thus, the carbon isotope data suggests hydrothermal graphite has been derived by mixing carbon from organic-rich pelites and calc-silicates of metasedimentary wall rocks (Rumble and Hoering, 1986).

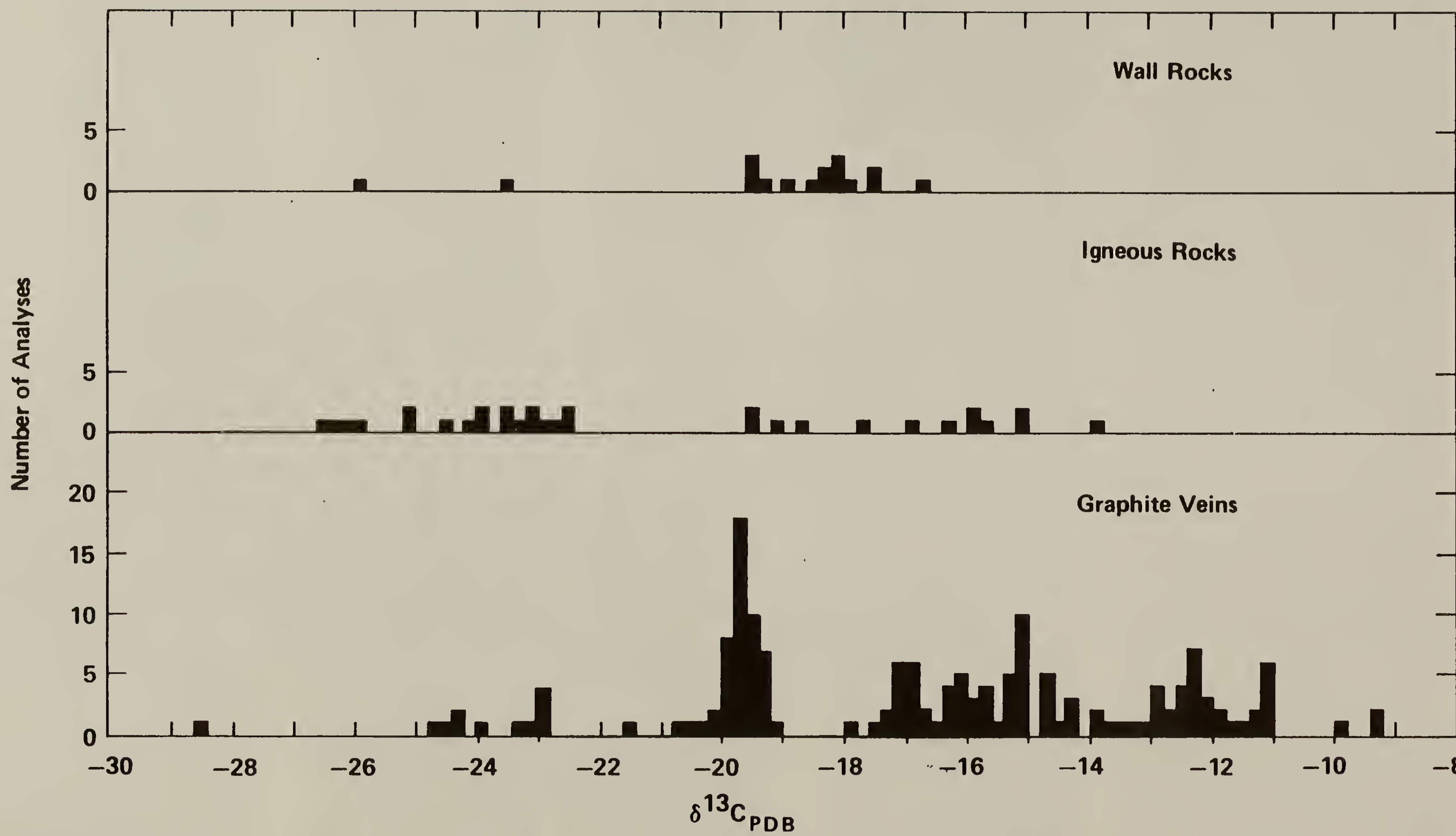


Figure 2 - Summary of d$^{13}$C values of graphite from wall rocks, plutons, and veins.

The oxygen isotope ratios of quartz from veins at the Bristol deposit range over the narrow interval of +13.9 to 14.6 $^{o}/oo$, $d^{18}O_{SMOW}$ The veins are surrounded by a halo about 2 Km in diameter throughout which wall rock gneisses are depleted in $^{18}0$ relative to their equivalents located at greater distances. Quartz separated from rocks inside the halo has $d^{18}O_{SMOW}$ values of +14.7 to 15.0 $^{o}/oo$. Rocks outside the halo contain quartz with values of +15.0 to 16.2 $^{o}/oo$ (Chamberlain and Rumble, in press). The vein values are consistent with the precipitation of quartz from fluids released by devolatilization reactions during metamorphism (Taylor, 1974).

## ORIGIN OF HYDROTHERMAL GRAPHITE

A theory of origin of graphite veins has been proposed by Rumble and Hoering (1986) that is consistent with field relations, petrology, stable isotope compositions, and phase equilibria. Carbon is mobilized by metamorphism of common sedimentary rocks. Metamorphism of shales containing reduced organic matter at low $f_{O_2}$ releases aqueous fluids with $CH_4 > CO_2$ and low $d^{13}C$ values. Metamorphism of argillaceous limestones drives off aqueous fluids with $CO_2 > CH_4$ and high $d^{13}C$ values. Carbon is transported as $CO_2$ or $CH_4$ in fluid-filled cracks that propagate through rocks by a mechanism of hydraulic fracture (Walther and Orville, 1982). Graphite precipitation occurs when two aqueous fluids with different $CO2/CH_4$ ratios are mixed (Rumble, et al., 1982). Intersection of propagating fractures with other fractures or with rocks undergoing devolatilization reactions results in fluid mixing. Note that fluid mixing not only leads to graphite precipitation but also mixes carbon from different sources resulting in graphite with intermediate $d^{13}C$ values.

Duke and Rumble (1986) proposed a theory of origin for spherulites in plutonic rocks that explains their association with retrogradation of igneous minerals. The pre- and syn-metamorphic plutons are infiltrated by metamorphic, C-O-H fluids along fractures and shear zones. The fluids are at or near the graphite saturation boundary. Hydration of primary biotite and feldspar, forming chlorite and sericite, causes dehydration of the fluids leaving remaining fluids enriched in $CO_2$ and $CH_4$. The compositions of residual fluids are driven across the saturation boundary and graphite precipitation ensues.

## OSGOOD VEIN DEPOSIT

The Osgood Mine in Nelson, N. H., is said to have furnished as much as 30 tons of plumbago in three months of mining (Jackson, 1844, p. 118). The presently exposed workings are the largest known in New Hampshire. Graphite ore occurs in a calc-silicate bearing, pelitic unit of the Silurian upper Rangeley Formation. The host rock is sulfidic and poorly bedded, consequently it is difficult to observe structures on the outcrops. We have had to rely on thin section study for much of our understanding of the deposit. Field trippers will, however, be able to study the graphite ore, thanks to the efforts of L. Baumgartner and W. Carey in clearing out more than a centuries' worth of debris.

The top of the graphite ore body shown in fig. 3 is an undulating surface, nearly planar, that strikes N-S and dips 40-45$^{o}$ westward. The ore is truncated along its northern boundary by a fault that strikes N63$^{o}$ to 74$^{o}$E and dips 45$^{o}$to 49$^{o}$NW. The ore body is variable in thickness, averaging 10 cm. The ore consists of graphite and ubiquitous but minor hercynite, with traces of ilmenite and pyrrhotite-chalcopyrite intergrowths.

Host rocks are rusty-weathered mica schists with calc-silicate nodules. The primary assemblage of pelites is cordierite-andalusite-biotite-garnet quartz. Porphyroblasts of cordierite, andalusite, and garnet are partially or wholly replaced by coarse sheaves of chlorite and radiating sprays of muscovite. Euhedral prisms of staurolite and needles of sillimanite usually occur in the chlorite. There are rare, small, corroded porphyroblasts cleaved like kyanite. The pelites also contain pyrrhotite, commonly more abundant than in the graphite ore, ilmenite, and flakes of micaceous graphite. In less aluminous mica schists, the primary assemblage quartz-plagioclase-biotite-muscovite-garnet is seen.

Calc-silicates occur as nodules, 20-40 cm in diameter, enclosed in mica schist. Primary assemblages of calc-silicates include quartz-plagioclase with various combinations such as amphibole-sphene-carbonate, amphibole-diopside-clinozoisite-sphene-carbonate, and amphibole-biotite-chlorite-tourmaline.

Graphite ore is always found to replace aluminous pelite. It has never been observed replacing calc-silicates

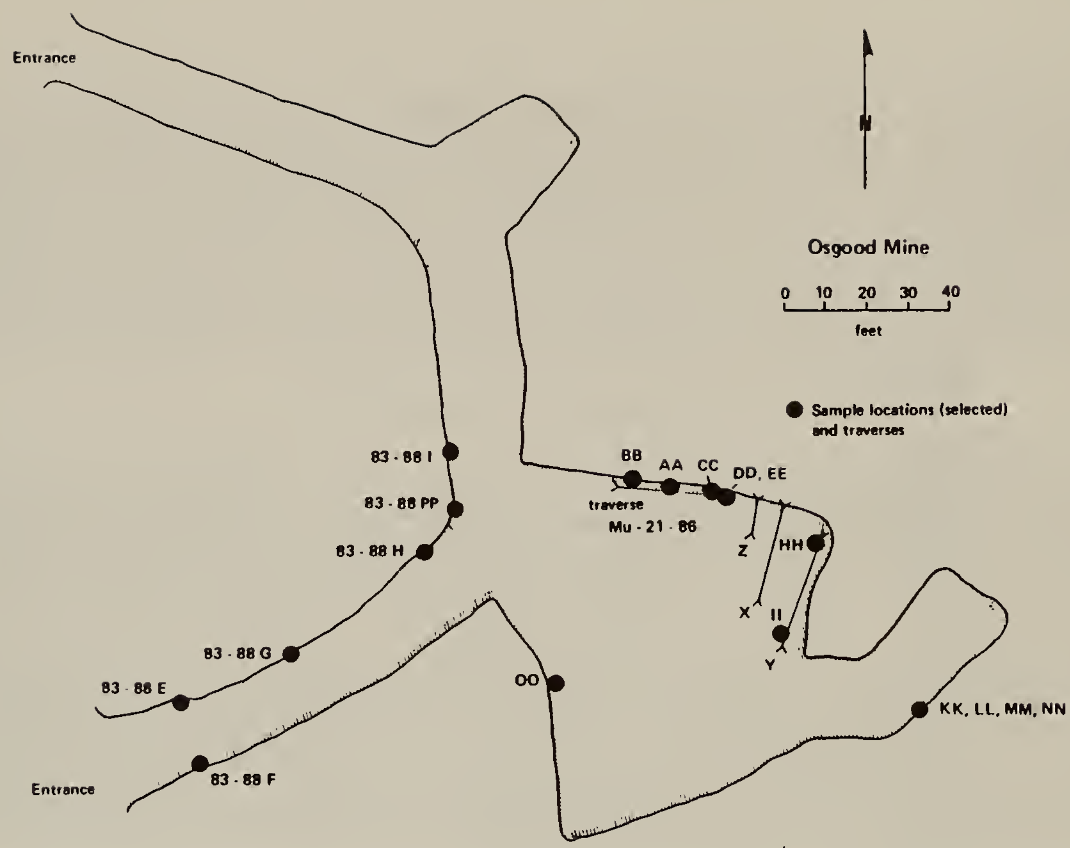


Figure 3a - Tape and compass map of Osgood mine.

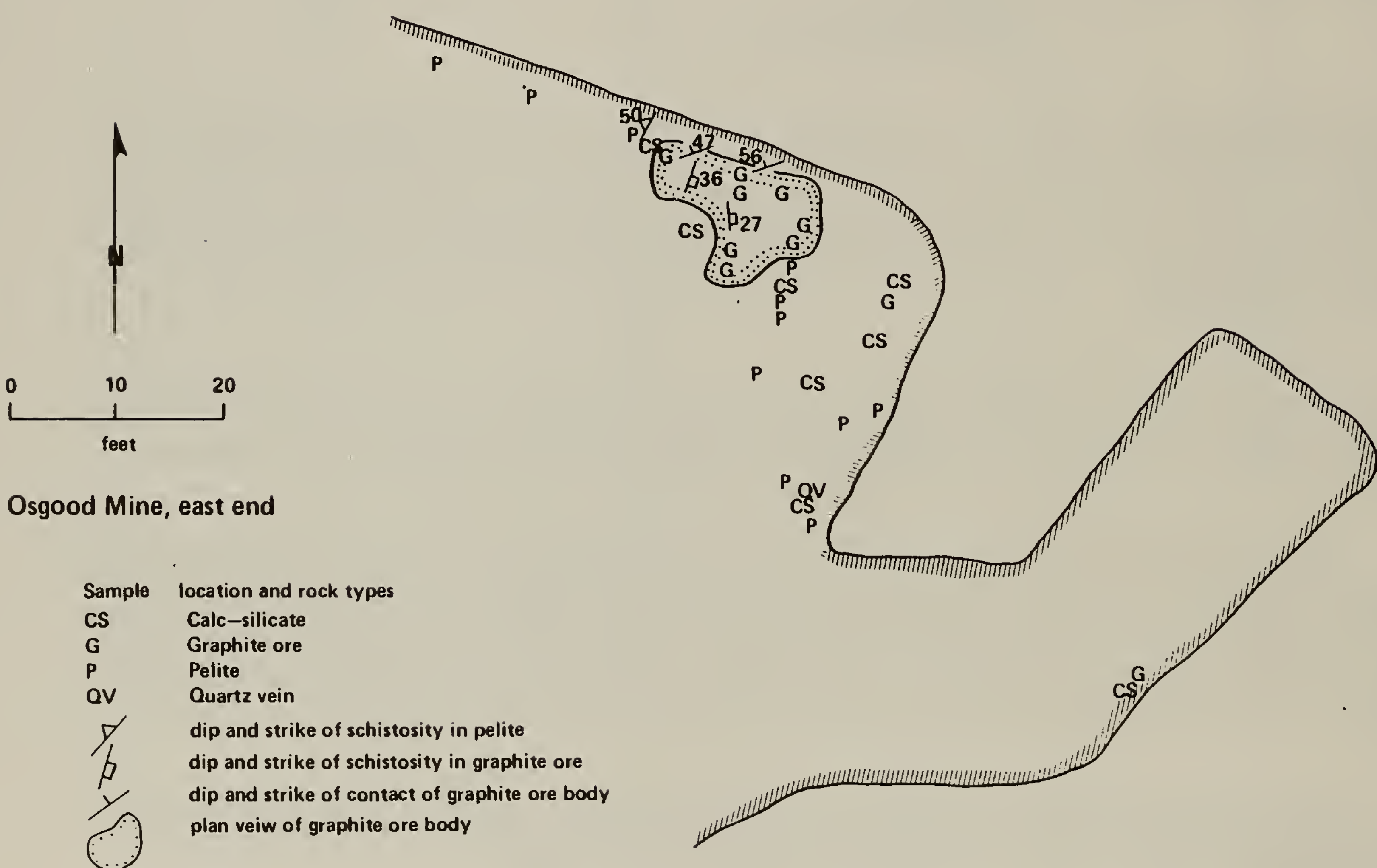


Figure 3b - Tape and compass map of Osgood mine, East end.

or mica schists that lack cordierite. The most massive ore consists of a schistose intergrowth of graphite flakes embedded with hercynite swirling around and completely enveloping relict porphyroblasts of cordierite and andalusite. Hercynite is evidently a key participant in the replacement reaction for it is not found outside ore zones. Retrogradation of relict porphyroblasts is always present in the ore. The breakdown of cordierite, andalusite, and biotite to more hydrous equivalents, however, occurs outside of ore zones, as well. It is not known why graphite mineralization is restricted to aluminous pelites. Perhaps graphitization is localized by the propensity of cordierite to undergo alteration. Perhaps there is a contrast in mechanical properties between aluminous pelites and other rock types that lead to fracturing and access of ore fluids. In contrast to Franklin Pierce and Mt. Kearsarge, there are no textures characteristic of mineral precipitation in open fractures.

Carbon isotope values of graphite ore range from -24.6 to -22.9 per mil (PDB) suggesting an affinity with sedimentary reduced organic matter (fig. 4).

## FRANKLIN PIERCE MINE

The mine was formerly owned by Franklin Pierce, 14th president of the United States (Hitchcock, 1878, p. 91). It was at one time the most profitable "plumbago" mine in the state, selling 20 tons of the mineral annually at a price of from 3 to 5 cents per pound (Jackson, 1844, p. 188). Eyewitness accounts dating from a time when the excavation was free of debris claim one of the graphite veins was as much as 14 inches thick (Nelson, 1957, p. 235). We have observed veins in broken blocks with a minimum thickness of 6 inches. Veins may be found in place where the fault intersects the aplite dike, in the widest part of the excavation (fig. 5).

The schists are part of a xenolith enclosed in an arm of the Mt. Cardigan pluton that extends SSW from the summit of Mt. Sunapee (Claude Dean's thesis) (fig. 1). The primary assemblage of the schists is biotite-sillimanite-muscovite-quartz-garnet-ilmenite-graphite with minor amounts of retrograde chlorite. Sulfide minerals are absent, in striking contrast to Osgood. Wall rock graphite is present as flakes that imitate the micaceous habits of muscovite and biotite.

Graphite ore occurs in veins, 1/4 to 6 inches thick, that crosscut bedding, schistosity, and an aplite dike. Vein minerals show textures that may indicate precipitation from fluids in open fractures. There are botryoidal, concentric intergrowths of graphite and quartz. One vein has bladed crystals of graphite, 3-4 mm long, projecting perpendicular from the wall of the vein into its center (fig. 6, left). Minerals co-precipitated with graphite in the veins are compatible with wall rock mineral assemblages. Two veins show euhedral sillimanite partly replaced by muscovite in a matrix of coarse, randomly oriented graphic flakes.

Examples of wall rock alteration and replacement are well developed at Franklin Pierce. In one specimen from a block, there is a replacement zone of graphite-sillimanite-ilmenite-quartz schist, 1 cm thick, separating massive, randomly oriented vein graphite from typical biotite-sillimanite schist. Schistosity in the replacement zone is continuous with schistosity in unaltered schist but is truncated by vein graphite. Another block shows a small vein, only 2 cm thick, surrounded by a wall rock replacement zone at least 6 cm thick (fig. 6, right). Fragments of relict wall rock minerals such as biotite, garnet, sillimanite, and muscovite, are completely enveloped by graphite and tourmaline in the replacement zone. The 1 cm tourmaline crystals are penetrated by graphite-filled cracks and contain inclusions of wall rock minerals. Note that wall rock alteration is not accompanied by substantial retrogradation of primary minerals as it is at Osgood. Primary minerals simply dissolve away.

Carbon isotope values of Franklin Pierce graphite are among the most enriched in $^{13}C$ that have been measured (fig. 4). The $d^{13}C$ values, moreover, occupy a much wider range, from -16 to -9 $^{o}/oo$, then any other deposit. Clues to an explanation for such an extreme case of carbon isotope disequilibrium may be found in fig. 6a & b. In both of the specimens illustrated, different textural varieties of graphite have contrasting $^{13}C/^{12}C$ ratios. Different $d^{13}C$ values are obtained from graphite formed at different times during mineralization. The isotopic composition of fluids that filled the fracture transecting the aplite dike with graphite evidently evolved continuously with time. Isotopic compositions may have changed in response to mixing different proportions of fluids from pelites and calc-silicates, to variations in $CO_2/CH_4$ ratios in vein fluids, or to changes in temperature. The sharp break in isotopic composition seen in fig. 6b, probably arose because the botryoidal graphite vein filing was deposited at a later time than graphite in the wall rock replacement zone, perhaps in a separate episode of mineralization.

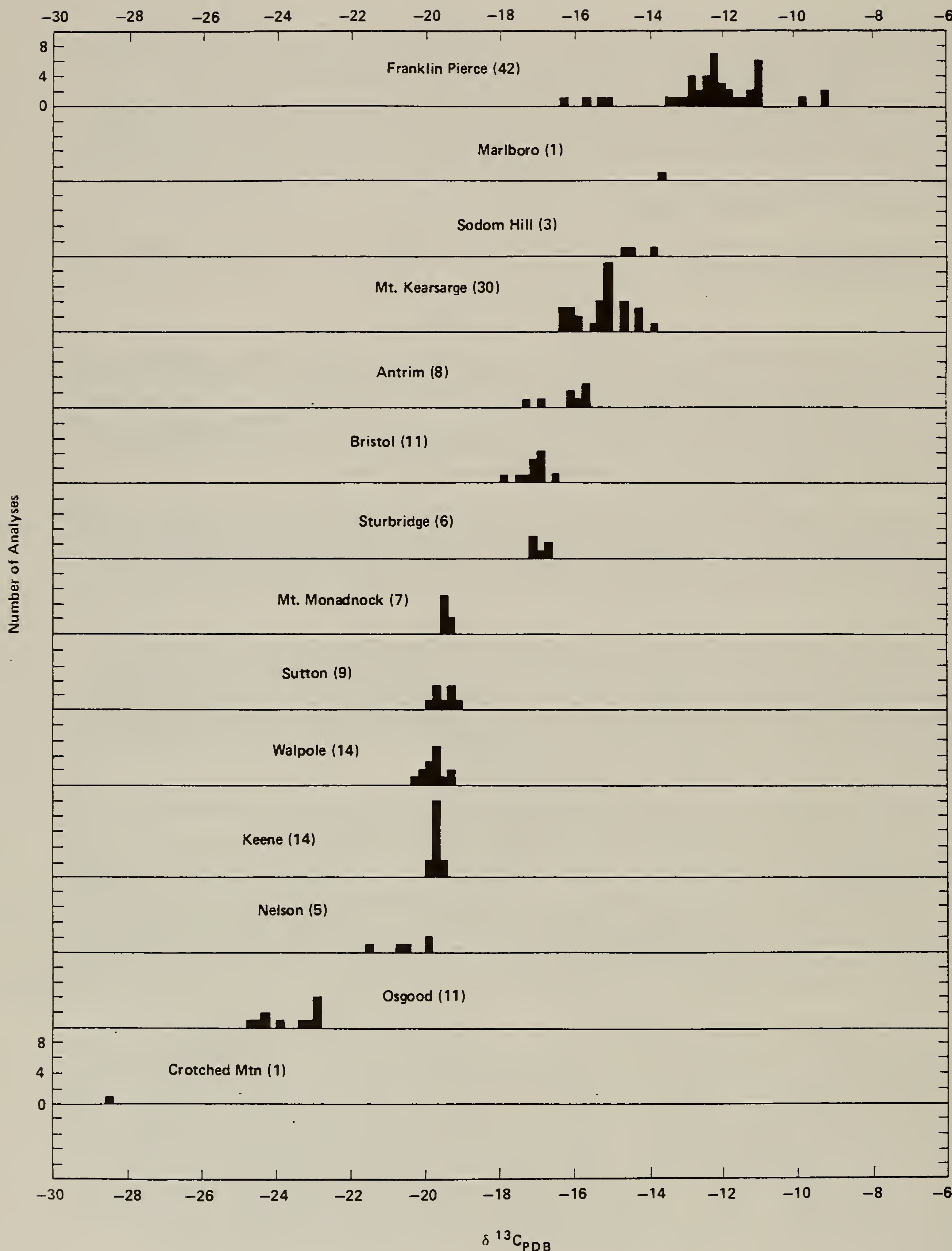


Figure 4 - Carbon isotope values of vein deposits. Number in parenthesis gives number of analyses

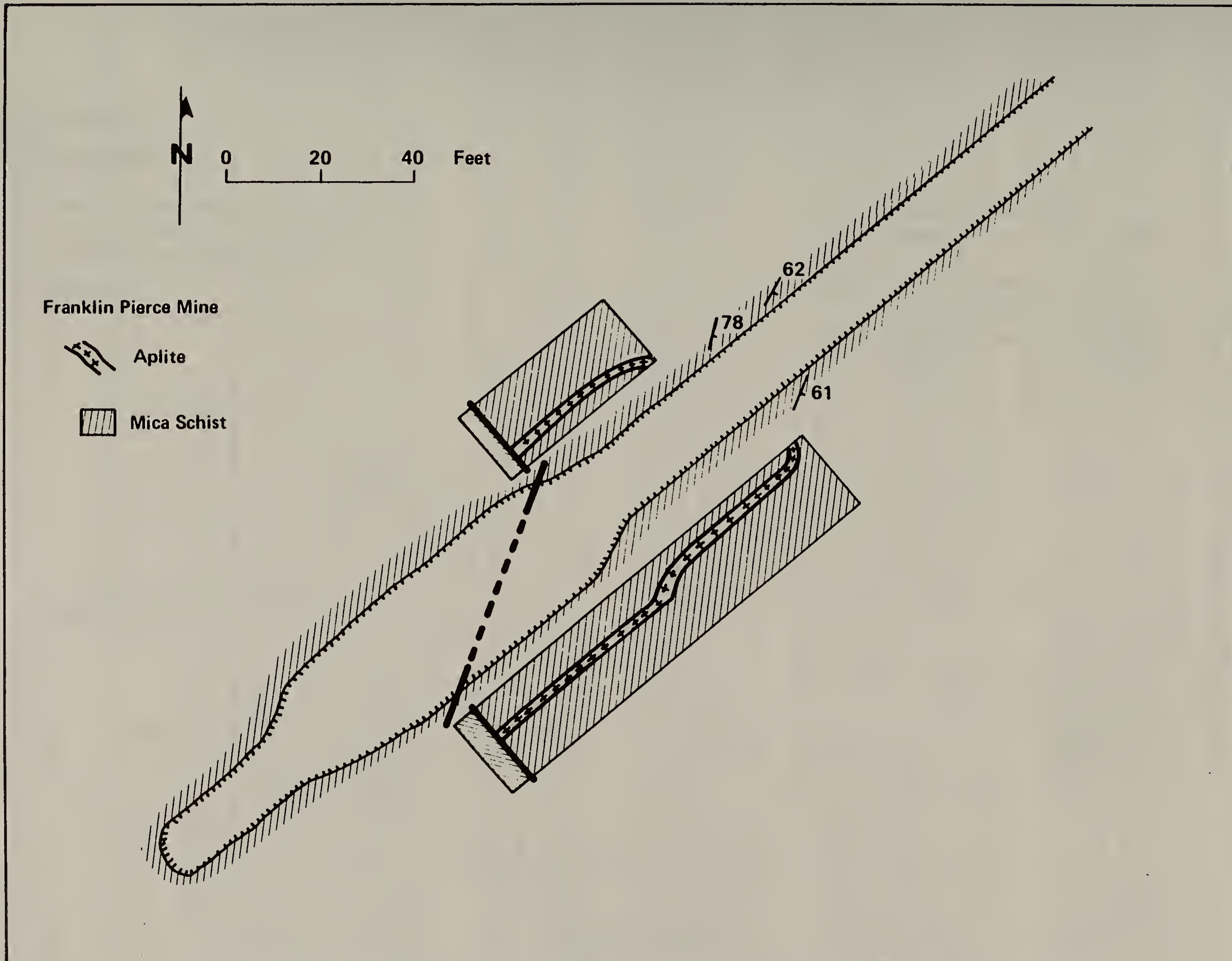


Figure 5 - Tape and compass map of Franklin Pierce Mine. Elevations of NW and SE walls of trench are shown.

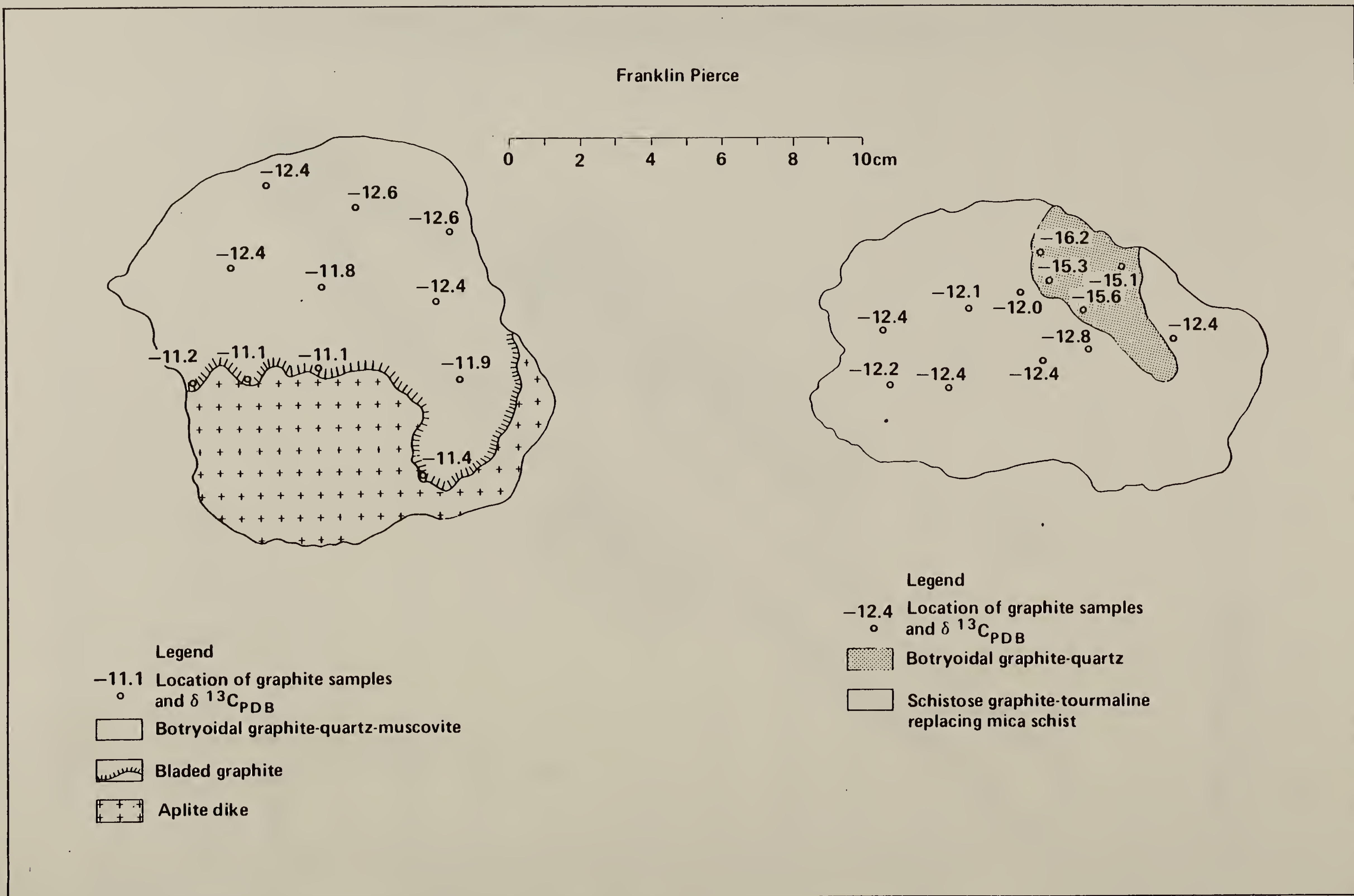


Figure 6a (left) and 6b (right) - Sketches of cut slabs of specimens from Franklin Pierce with $d^{13}C$ values.

## MT. KEARSARGE MINE

The Mt. Kearsage mine was worked and abandoned prior to publication of the Salisbury town history in 1890. Despite the obscurity of its history, the deposit is among the most fascinating to visit for study. Graphite veins and nodules can be found in place. Their geological relationships are well-exposed.

Graphite veins and nodules appear in an aplitic phase of the Spaulding Subseries (fig. 1) near its contact with the Lower Devonian Littleton Formation (fig. 7). Xenoliths of mica schist are present in the margin of the pluton. Unaltered samples of the Spaulding consist of perthitic alkali feldspar, quartz, plagioclase, biotite, garnet, muscovite, and ilmenite. The primary assemblage in Littleton schists is quartz-plagioclase-muscovite-biotite-prismatic sillimanite-garnet-ilmenite-micaceous graphite-rutile. An alteration zone extends 10-20 m. on either side of the contact. Within the zone primary silicate minerals of both the pluton and its wall rocks are converted to chlorite and sericite and the rocks are graphitized.

The vein labeled "BGV" botryoidal graphite vein, on fig. 7, has spectacular hydrothermal textures. The vein is an example of hydrothermal brecciation, a process often observed in epigenetic veins. An early generation of randomly oriented, mm-size graphite flakes was broken into fragments of about 1/2 cm in size during hydraulic fracturing. As the graphite "clasts" were tumbled in fresh aliquots of hydrothermal fluids, layer upon layer of finer-grained graphite was deposited concentrically around them. Quartz, muscovite, and chlorite were co-precipitated with graphite at this time. Some samples show xenocrysts of garnet from wall rocks that also were enveloped by graphite. Following this turbulent stage of deposition, there was a final, minor phase of fracturing during which graphite and ilmenite veins 1 mm thick crosscut the botryoidae. The resulting texture is distinctive in hand specimen: one can see 1/2 cm graphite balls and pop them out of the vein with a knife blade.

The wall rock of BGV is intensely altered. It resembles the plutonic Spaulding more closely than rocks of the Littleton Formation. In view of the variety of hydrothermal events recorded in BGV, it is not surprising that its wall rocks should display superimposed alteration textures. The most obvious alteration is chloritization and sericitization of primary biotite, garnet, and feldspar. But, there are also puzzling occurrences of staurolite and sillimanite adjacent to the vein. Perhaps the aluminous minerals resulted from an early episode of wall rock alteration analogous to that at Franklin Pierce. Or, perhaps the border of the pluton was contaminated by wall rocks.

Graphitization of wall rocks at BGV is of special interest because it sheds light on the origin of spherulitic graphite, mentioned above as pervasively distributed in pre- and syn-metamorphic plutonic rocks. Both plutonic rocks and the schist xenoliths at BGV contain spherulites, 5-10 mm in diameter, that show a "maltese cross" extinction pattern under crossed nicols. Spherulites are not seen in these rock types outside the alteration zone along the pluton-mica schist contact. The similarity of these spherulites to those found in the Mt. Cardigan and Mt. Kinsman plutons reinforces the interpretation ascribing the latter occurrences to a hydrothermal origin.

Another equally remarkable hydrothermal texture can be found at the locality labeled "NG", nodular graphite (fig. 7). Here nodules of graphite have textures indicating replacement of the Spaulding. The remains of a nodule 30 cm in diameter are still exposed but most of them are a few cm in size (fig. 8). The nodules display spectacular clusters of mm-long graphite crystals radiating from a common center (see cover photo of Geology, June, 1986). These "sunbursts" of graphite crosscut primary minerals of the plutonic rocks. A less common texture has microscopic concentric layers of alternating, very fine-grained graphite and silicates. The microstructures resemble a reaction front of solution followed by precipitation propagating through the aplite, rather like Liesegang rings. Polished sections of larger nodules found in the dump suggest that they were formed by continued growth and coalescence of radiating clusters and concentric layers of graphite. The nodules, however, were not disrupted by hydraulic fracturing as at BGV. The Spaulding hosting the nodules has serecitized feldspar and chloritized biotite but alteration is less intense than at BGV. There is no staurolite or sillimanite in the wall rock of the nodules.

Graphite mineralization was not limited to the pluton as is suggested by currently known outcrops. A search of blocks in the dump revealed a graphite vein with hydrothermal textures crosscutting mica schist with prismatic sillimanite.

Carbon isotope values at Mt. Kearsarge overlap the lower range of those measured at Franklin Pierce and are higher by 6‰ than those at Osgood (fig. 4). The variation in $d^{13}C$ between adjacent nodules at NG is as much as 2 ‰ over distances of a few cm (fig. 8). The explanation of localized carbon isotope heterogeneity at Mt.

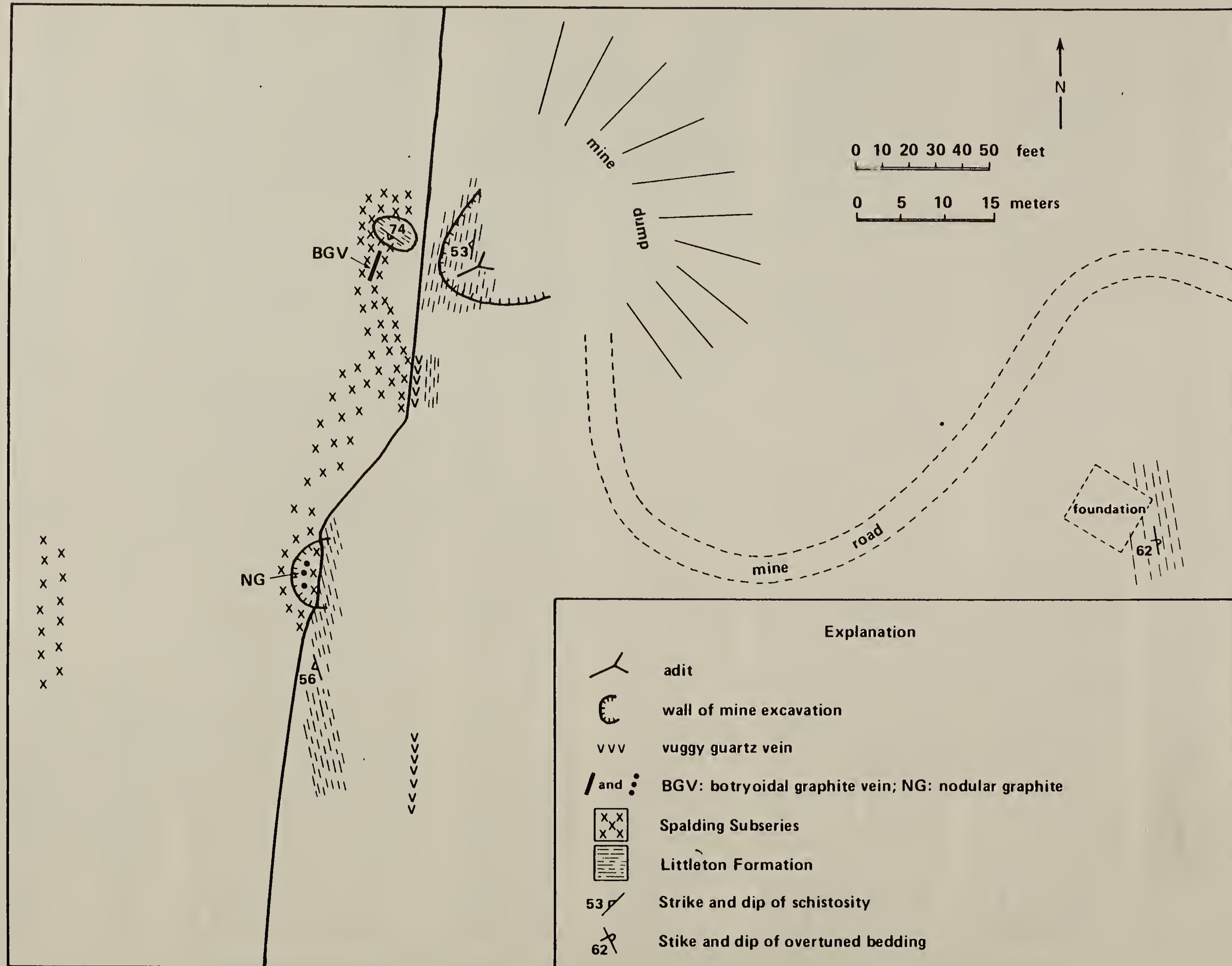


Figure 7 - Tape and compass map of Mt. Kearsarge mine.

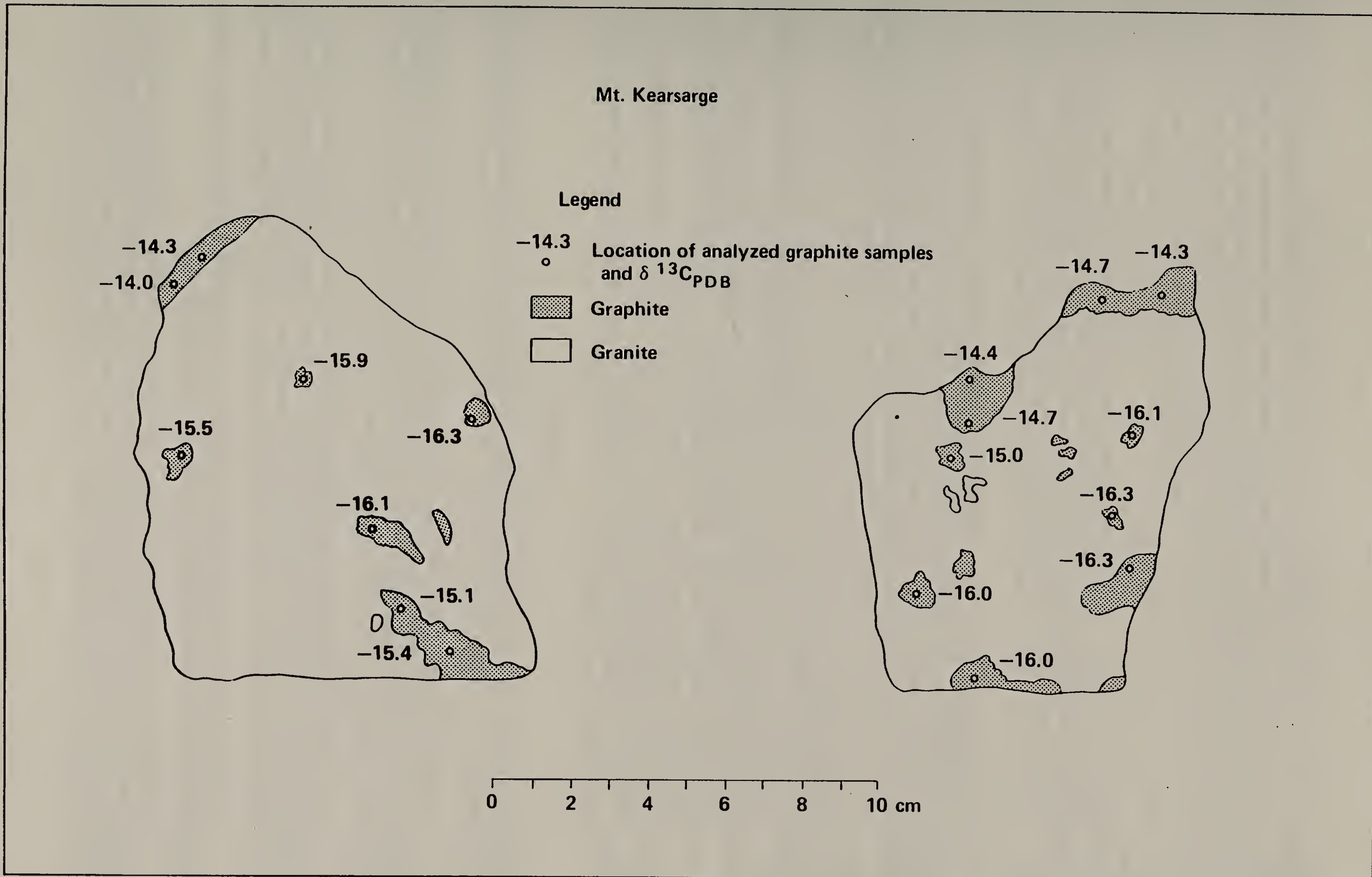


Figure 8 - Sketches of cut slabs of specimens from Mt. Kearsarge mine with $d^{13}C$ values.

Kearsarge is much the same as for Franklin Pierce. Textures give evidence of successive episodes of graphite mineralization. The isotopic composition of graphite precipitated from fluids is controlled by the bulk $d^{13}C$ of the fluid, by its $CO_2/CH_4$ ratio and by temperature. Changes in either or all of these factors could have led to differences in $d^{13}C$ from one generation of graphite to the next.

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## ROAD LOG

Assembly point: Keene State College Commons Parking: 8 am, Saturday, Oct. 15: Quandrangles: Osgood Mine - Mt. Monadnock 15'; Franklin Pierce Mine - Sunapee 15'; Mt. Kearsarge Mine - Mt. Kearsarge 15'

1. Proceed northward on N. H. Rte 9 (direction towards Concord) for about 9 miles and turn right (East) on a paved road with a signpost marked "Nelson". You will know if you have gone too far if you reach Munsonville on Granite Lake. The correct turn off Rte 9 is 1-1/2 miles south of Granite Lake. Proceed East on paved road 2 miles to Nelson Village. We will pause to wait for stragglers in the center of the village, beside the Green.

2. Go straight to the East on gravel road from the Nelson Village Green. Do not continue to the right, downhill on the paved road. Proceed East, passing town barn, then curve Northeast on gravel road. Park cars along the road 1-1/2 miles from the Green. Osgood Mine is a few hundred yards to SE, at the end of an overgrown haulage road. The mine is located on the lower, western slope of Osgood Hill (N. central 9th of Monadnock 15' quadrangle.

3. The Osgood Mine has many unstable rubble blocks. It is also thickly overgrown. Please be careful of your footing and do not allow tree branches to slap your fellow field-trippers.

4. Return to N. H. Rte 9 via Nelson Village. Turn right on Rte 9, Proceed about 8 miles NE on Rte 9, passing through village of South Stoddard. Turn left on N. H. Rte 123. Proceed NW on Rte 123 through villages of Mill Village and Stoddard to N. H. Rte 10 (about 8 miles). Turn right on N. H. Rte 10. Drive N. on N. H. Rte 10 for about 15 miles to intersection of Rte 10 with N. H. Rte 31. Turn hard right (acute angle) on to N. H. Rte

31. Drive S. on Rte 31 for 1.6 miles to "Mountain Road". Turn left on Mountain Road. We will wait for stragglers at this turn. Proceed NE on Mountain Road for about 1 mile and park. Walk down steep hill, cross bridge over brook, go past stone foundation, and turn right at sign that says "Pillsbury State Park, Washington". Then after 100 ft., turn right at yellow tape flags onto lumber trail. The woods have been recently logged over so the old, orange-blazed trail has been torn up. A route to the Franklin Pierce Mine was flagged with yellow tape in 1987. If there has been no additional logging and if the flags are still in place we will have no trouble finding the mine. Please do not charge off into the woods without checking with the field trip leaders. The Franklin Pierce mine is a long, narrow trench. Please do not climb on the walls of the trench. The floor of the trench at the far, SW end of the trench rises rather steeply. The rubble here is unstable. Please do not kick blocks down on your fellow field trippers.

5. Return to intersection of N. H. Rte 31 and Rte 10 (turn right, to N., on Rte 31). Continue N. on Rte 31 for 3/4 mile to Goshen Village. Turn right on Rand Pond Road. Drive NE about 5 miles, past Rand Pond, to N. H. Rte 103. Turn right on N. H. Rte 103, towards Mt. Sunapee State Park. Drive E. on Rte 103 for 18.5 miles to village of Warner. We will wait for stragglers along Rte 103 in Warner.

6. Proceed N. from Warner on Tucker Pond Road (beginning is paved but turns to gravel.) Drive about 6 miles, around Tucker Pond, to Smith's Corners. Turn left at Smith's corners Proceed 1-1/4 miles N. of Smith's corners to Scribners Corners. Turn left at Scribners corners and drive WSW on old road. The road is no longer maintained and bridges may have washed out. We will drive as far as we can and then walk to Mt. Kearsarge mine. Be careful driving up over eroded parts of road where ledges are exposed. Low slung cars are liable to bottom out. Check bridges before driving on them.

7. The Mt. Kearsarge mine is on the E. slope of Mt. Kearsarge. Please do not go in the adit; there is plenty to see above ground. The steep slope above the adit is littered with loose blocks. Please watch your step and do not dislodge them onto fellow field trippers.

8. Return to Keene via Warner, Rte 103, Rand Pond Road, and Rte 10. An alternate return route is to take Interstate 89 South from Warner to Rte 9. Then follow Rte 9 all the way to Keene.

# THE CONNECTICUT VALLEY-GASPÉ SYNCLINORIUM IN SOUTHEASTERNMOST VERMONT

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## INTRODUCTION

In southeastern Vermont the regionally extensive Connecticut Valley-Gaspé synclinorium separates the Bronson Hill anticlinorium, with its Oliverian gneiss-cored domes, to the east from the Green Mountain anticlinorium to the west (Doll and others, 1961). This synclinorium is principally underlain by a thick sequence of late Ordovician, Silurian and early Devonian flysch and calcareous flysch and minor, but important, interbedded mafic volcanics. These rocks have undergone variable degrees of metamorphism (chlorite to kyanite zones) and have been multiply folded and intruded during the Acadian Orogeny. A belt of domes, occurring in the Connecticut Valley-Gaspé synclinorium, extends southward from east-central Vermont to Connecticut, west of the Connecticut River. They are somewhat analogous to but more widely spaced than the domes of the Bronson Hill anticlinorium to the east. Large recumbent folds are found in the strata mantling these domes in eastern Vermont (Doll and others, 1961). The Standing Pond Volcanics is an important marker unit outlining many of these recumbent folds and domes. The axial surfaces of the recumbent folds have been arched by the later doming. The arcuate, closed, double band of the Standing Pond Volcanics around the southern end of the Guilford done in southeastern Vermont (fig. 1) outlines such a refolded recumbent fold. One of the purposes of this field trip is to investigate this fold and the proposed east-facing recumbent anticline above it. Other stops will be made to view the Black Mountain Granite, an important key to determining the time of deformation; the Waits River Formation in the exposed core of the Guilford dome; and the contact between the Putney Volcanics and the Littleton Formation, which separates the "Vermont" and "New Hampshire" sequences (Billings, 1956).

This field excursion is an updated version of one led during the 1972 N.E.I.G.C. (Hepburn, 1972). For a more complete description of the geology of the Brattleboro area than can be presented here, detailed geological maps, etc., the reader is referred to Hepburn and others, 1984.

## STRATIGRAPHY

Brief descriptions of stratigraphical units pertinent to this excursion are summarized below. For more detailed descriptions of the stratigraphy of this area please refer to Doll and others (1961), Chang and others (1965), and Hepburn and others (1984), and references therein.

The Standing Pond Volcanics forms a distinctive marker formation for mapping and by defination separates the Waits River and Gile Mountain Formations. However, it should be recognized that in southeastern Vermont the Standing Pond Volcanics does not everywhere exactly separate calcareous rock types (characteristic of the Waits River Formation) from non-calcareous rocks (characteristic of the Gile Mountain Formation). Calcareous rocks in the Gile Mountain Formation similar to those in the Waits River Formation have been designated the marble member of the Gile Mountain (fig. 1, Stop 9). The quartzitic member of the Waits River Formation contains schistose and feldspathic quartzites and mica schists, similar to those in the Gile Mountain Formation.

The Putney Volcanics (Stops 1 and 2) consists of a belt of rocks that were formerly included in the Standing Pond Volcanics (Doll and others, 1961) but now is designated as a separate formation (Trask, 1980).

Determination of the stratigraphic order of the rocks in the "Vermont" sequence has been a long standing problem. No definitive evidence for the facing of the Waits River, Standing Pond, and Gile Mountain formations has yet been found in southeastern Vermont. However, Fisher and Karabinos (1980) reported good stratigraphic topping evidence near Royalton, Vermont, that indicates the Gile Mountain is younger than the Waits River Formation. Thus, the sequence, oldest to youngest, of Waits River, Standing Pond, and Gile Mountain, as shown on figure 1 is favored, although other possibilities can not be ruled out.

Good stratigraphic topping evidence has been found in the Brattleboro quadrangle at several sites within the Putney Volcanics and at its contact with the Littleton Formation (Hepburn and others, 1984). This evidence based on small cross-bedded sequences, consistantly indicates that the stratigraphy gets younger to the east across this contact, i.e. that the Littleton overlies the Putney (Stop 2). This implies the package of rocks between the Shaw Mountain Formation and the Putney Volcanics, the "Vermont" sequence, is pre-Littleton (Early Devonian) in age. Recently, Bothner and Finney (1986) reported the recovery of Middle to Upper Ordovician graptolites from the Waits River Formation near Montpelier, Vermont. Thus, the age range of the "Vermont" sequence likely spans the time from middle or late Ordovician to earliest Devonian. However, because temporal boundaries have not yet been established for this sequence of rocks, they are herein (fig. 1) labeled Silurian, conforming with Hepburn and others (1984).

Stratigraphic units in southeastern Vermont (Brattleboro quadrangle) include:

MIDDLE ORDOVICIAN

**Barnard Volcanic Member, Missisquoi Formation:** Massive to schistose porphyritic and non-porphyritic amphibolites, feldspar-rich gneisses, and layered gniesses. Thickness, 4000-8000 feet.

**Cram Hill Member, Missisquoi Formaton:** (Following Doll and others, 1961.) Rusty-weathering, black carbonaceous phyllite and schist with interlayered amphibolite. Thickness, 2000-4000 feet.

SILURIAN (May include in part, middle to late Ordovician and/or early Devonian)

**Shaw Mountain Formation:** (Russell Mountain Formation of Hepburn and others, 1984) Quartzite and quartz-pebble conglomerate, hornblende fasciculite schist, amphibolite, and mica schist. Thickness, 0-20 feet.

**Northfield Formation:** Gray mica schist with abundant almandine porphyroblasts, minor impure quartzite and punky-brown weathering impure marble interbeds. Thickness,1000-2500 feet.

**Waits River Formation:** Mica schist (phyllite at lower metamorphic grades) and calcareous mica schist with abundant distinctive interbeds of punky-brown weathering impure marble; thininterbeds of impure quartzite. **Quartzitic member:** feldspathic and micaceous quartzite interlayered with mica schist. Formation thickness, 3000-7500 feet.

**Standing Pond Volcanics:** Medium-grained amphibolite and epidote amphibolite; garnet-hornblende fasciculite schist. **Eastern band:** plagioclase-biotite-hornblende-quartz granulite and gneiss. Formation thickness, 0-500 feet.

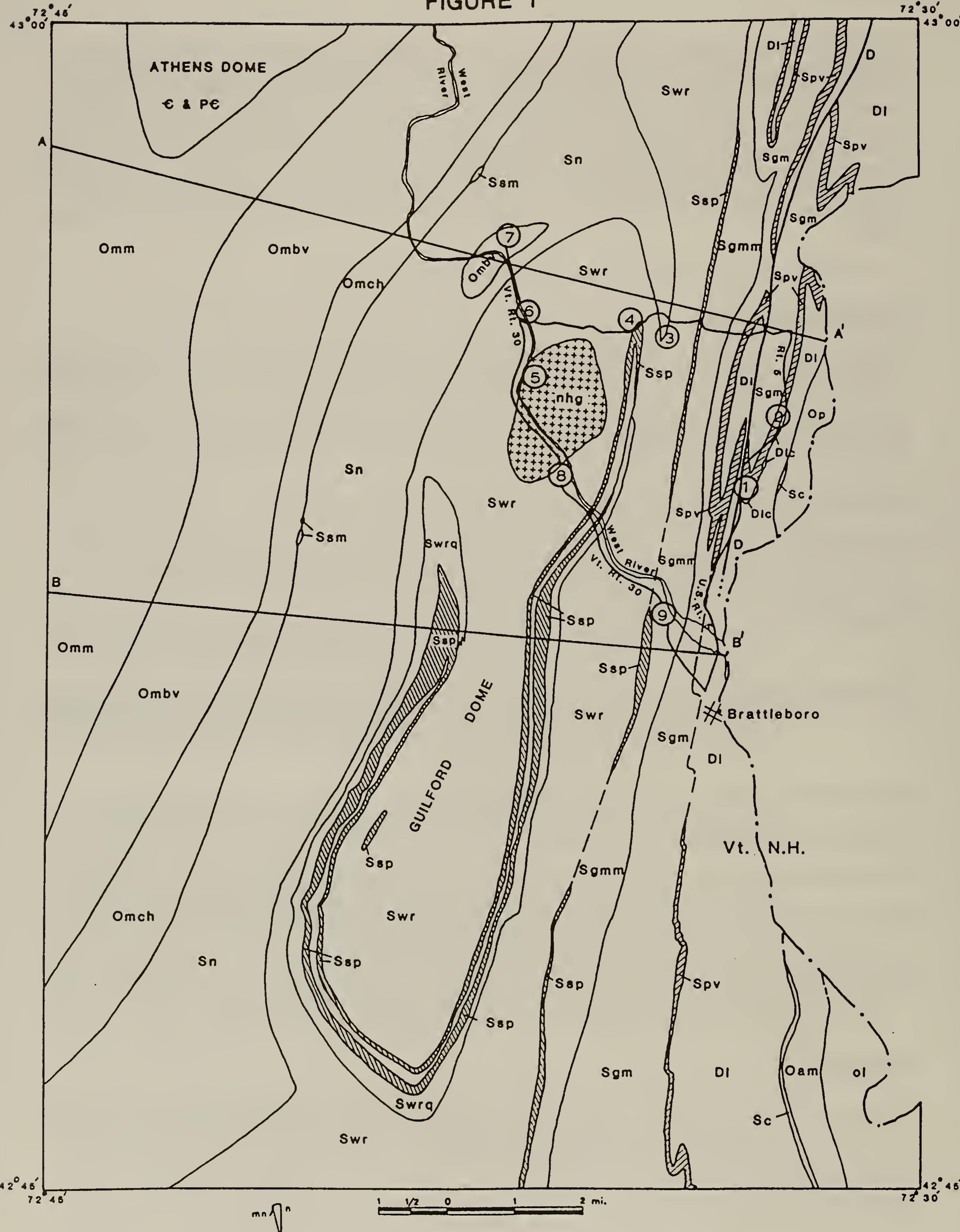
**Gile Mountain Formation:** Light gray to gray, micaceous and feldspathic quartzite and mica schist; gray phyllite and slate with interbedded, thin micaceous quartzite and rare impure marble. **Marble member:** gray to black phyllite with interbeds of punky-brown weathering, impure marble and micaceous quartzite. Formation thickness, 2500-5000 feet.

**Putney Volcanics:** Light, greenish gray phyllite; buff to light brown weathering feldspathic phyllite; thin beds of feldspathic granulite; and minor gray slate. Thickness, 0-400 feet.

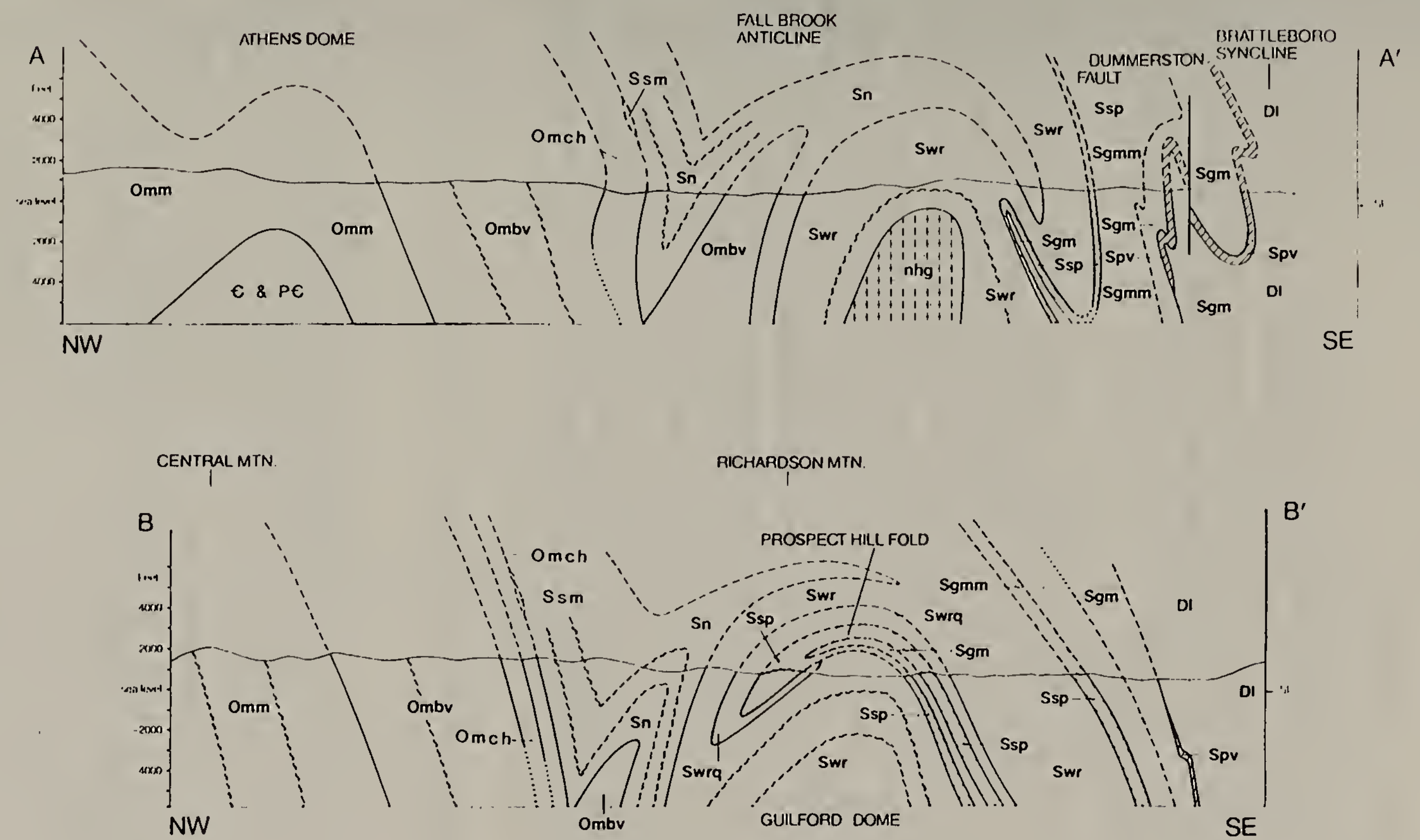
DEVONIAN

**Littleton Formation:** Gray slate or phyllite with interbeds of quartzite. **Conglomeratic member** (at base of formation): Lenses of polymict conglomerate with a gray slate matrix; pebbles abundant to scarce. Formation thickness, 5000-6000 feet.

FIGURE 1




GEOLOGIC MAP OF THE GUILFORD DOME AREA, SE VERMONT




LEGEND FIGURE 1

METAMORPHIC ROCKS

Dl LITTLETON FM.

Dlc LITTLETON FM.,congl. mbr.

Spv PUTNEY VOLCANICS

Sgm GILE MOUNTAIN FM.

Sgmm GILE MT. FM., marble mbr.

Ssp STANDING POND VOLC.

Swrq WAITS RIVER FM., qtzt. mbr.

Swr WAITS RIVER FM.

Sn NORTHFIELD FM.

Ssm SHAW MOUNTAIN FM.

Omch MISSISQUOI FM., Cram Hill Mbr.

Ombv MISSISQUOI FM., Barnard Volc. Mbr.

Omm MISSISQUOI FM., Moretown Mbr.

Sc CLOUGH QUARTZITE

Op PARTRIDGE FM.

Oam AMMONOOSUC VOLC.

PLUTONIC ROCKS

nhg NEW HAMPSHIRE PLUTONIC SERIES

Black Mt. Granite

ol OLIVERIAN PLUTONIC SERIES

D Fault, D on Downthrown side

3 Field Trip Route & Stop

Geology and Cross Sections Modified after

Hepburn, Trask, Rosenfeld, and Thompson, 1984

## INTRUSIVE ROCKS

**Black Mountain Granite:** Medium-grained garnet bearing two mica granite to granodiorite. Hayward and others (1988) recently determined a Carboniferous crystallization age of 326 ± 17 Ma (Rb-Sr whole rock) for the Black Mountain Granite. Shields (1977, see also Hepburn and others, 1984) gravity study of the Black Mountain Granite indicates the body extends for 5 km below the current surface and was intruded more or less along the steeply west dipping axial surface of the Guilford dome with a nearly vertical eastern contact and a western contact dipping approximately 55°.

## STRUCTURAL GEOLOGY

The major tectonic features in the Connecticut Valley-Gaspé synclinorium in southeastern Vermont formed during the Acadian Orogeny, between the end of sedimentation in the early to mid-Devonian and the crystallization of the Black Mountain Granite. Late normal faulting and possibly some minor folding occurred during the Mesozoic. Two major stages of folding dominate the structural evolution of the area, these are: (1) the development of large recumbent folds, followed by (2) the rise of domes. On the excursion we will see the Prospect Hill recumbent fold and the Guilford dome as examples of these major folding stages.

The doubly-closed loop of Standing Pond Volcanics around the southern part of the Guilford dome outlines the Prospect Hill recumbent fold, named for exposures at the hinge (Stop 4). The Gile Mountain Formation forms the core of the fold. Originally the Prospect Hill fold had a subhorizontal axial surface and a hinge striking northeast-southwest. The subsequent doming about a roughly N-S axis arched the axial surface of the recumbent fold, so that now the hinge plunges moderately northeast and southwest away from the axial trace of the Guilford dome. An early, tight, now overturned, steeply east-dipping synform must lie between the Standing Pond bands in the doubly-closed loop and a third band lying to the east of the Guilford dome (fig. 1). The hinge line where the Standing Pond rocks cross the axial surface of this synform is not seen in the Brattleboro area and is presumably buried. This synform, the Northfield Formation around the northern end of the Guilford dome, and the Fall Brook anticline which exposes the Barnard Volcanics, are interpreted as the upper (anticlinal) portion of the Prospect Hill fold (fig. 1, Cross-section A). Figure 3 shows in schematic form the assumed evolutionary stages in the development of the Prospect Hill fold and Guilford dome and the relationship of the metamorphism to the major folding stages. Note that Rosenfeld (1968, and in Hepburn and others, 1984, p.93-100), from studies of rotated garnets and using a different stratigraphical order than is assumed here, produces the Prospect Hill fold by a somewhat different sequence of events, involving initial westward transport and backfolding with intrastratial flow accompanying doming.

It is very likely that the Prospect Hill fold is continuous with the Ascutney sigmoid in the Saxtons River quadrangle to the north (Rosenfeld, 1968; Doll and others, 1961).

The Guilford dome, which occupies much of the central portion of the Brattleboro quadrangle (fig. 1), is a large, elliptical, doubly-plunging anticline formed during the second major stage of deformation. The Waits River Formation forms the exposed core of the dome. The foliation dips away in all directions from the axial trace, which strikes slightly east of north and plunges moderately to the north and south at the ends of the anticline. The axial surface of the dome dips very steeply to the west. A small depression in the exposed central portion of the dome divides it into a northern and southern lobe. The axial trace of the dome lies closer to its eastern side. Here, the foliation has steep dips a short distance east of the axial trace. Dips are more gentle to the west. Bedding, with a schistosity parallel to it, has been arched by the dome.

### MINOR FOLDS

Minor folds of at least five different stages are present in Connecticut Valley-Gaspé synclinorium in the Brattleboro area (Hepburn, 1975). These stages of minor folding include:

$F_1$ Small isoclinal folds in layering, with schistosity developed parallel to the axial surfaces (Stop 4).

$F_2$ Tight to isoclinal folds congruous with the large-scale recumbent folding (Prospect Hill fold). These fold the schistosity and the $F_1$ folds and plunge moderately NE or SW. Weak to moderate axial-planar cleavage (Stop 4).

F3 Open folds, particularly west and south of the Guilford dome. Excellent slip-cleavage developed parallel to the axial surfaces. The axial surfaces generally strike NE and dip steeply NW. The hinges plunge moderately NE. Excellent crinkle lineations occur at the intersection of this slip-cleavage and the schistosity surfaces in the pelitic rocks.

F4 Open folds, buckles or warps in the foliation that are of one or more generations and fold the slip-cleavage.

F5 Large open folds found only in the eastern part of the area (fig. 1) that offset the Putney Volcanics with an east-side-north movement. Plunge is moderately to steeply north. Kink bands also found along the eastern part of figure 1 are the youngest minor folds and may be related to the above F5 folds or may be younger.

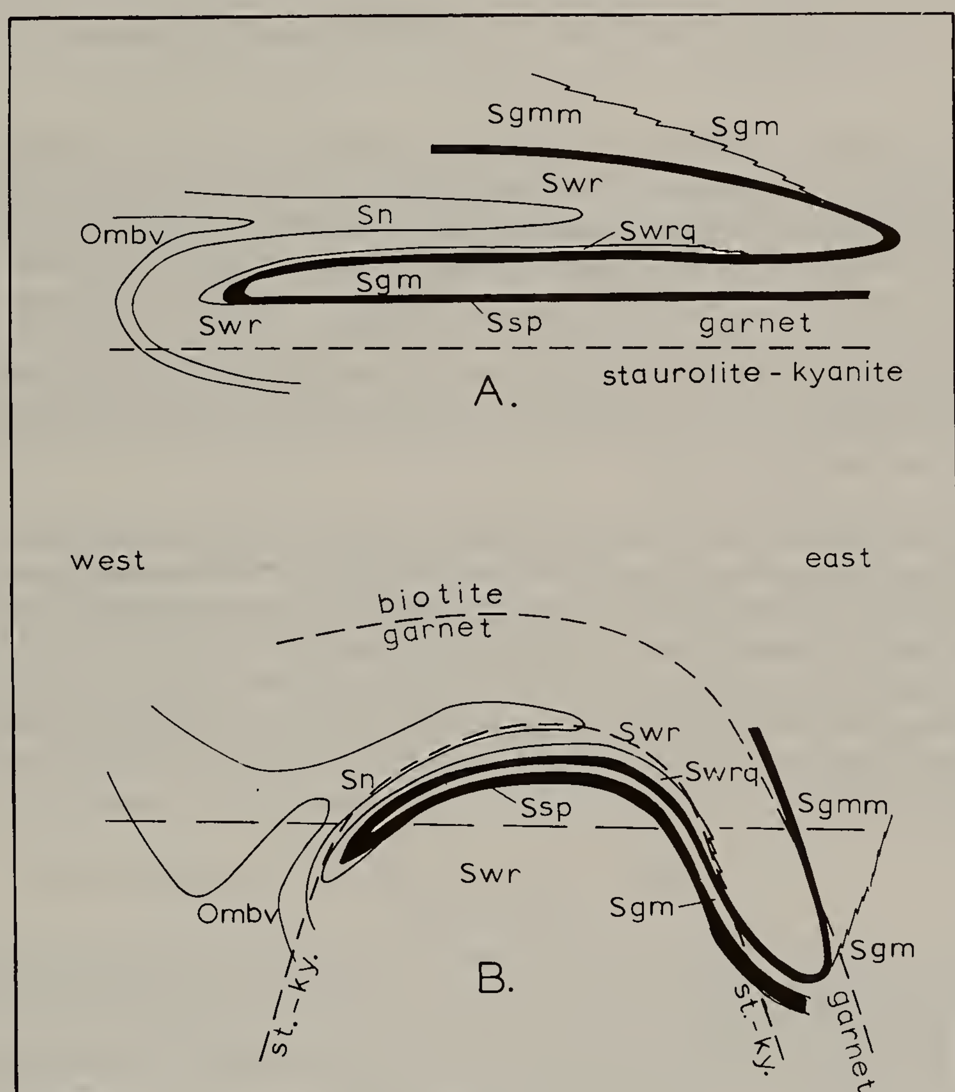



**Figure 3.** Schematic cross sections showing the evolution of the structural features in the two major stages of deformation in the Guilford dome area. The Standing Pond Volcanics are shown in black. **(A).** The Prospect Hill fold at the end of the first major stage of deformation, before the rise of the Guilford dome. The dashed line represents a hypothetical staurolite or kyanite isograd. **(B).** Prospect Hill fold following the second major stage of deformation, after the rise of the dome. Horizontal line represents the present erosion surface. The dashed lines show the assumed present distribution of isograds. Modified after Hepburn and other (1984). Formations: Sgm=Gile Mt. Fm.; Sgmm=Gile Mt. Fm., marble mbr.; Ssp=Standing Pond Volcanics; Swr=Waits River Fm.; Swrq=Waits River Fm., quartzitic mbr.; Sn=Northfield Fm.; Ombv= Missisquoi Fm., Barnard Volcanic Mbr.

## METAMORPHISM

A belt of low-grade metamorphic rocks (chlorite zone) occurs in the eastern part of the area and roughly follows the Connecticut River. This low is of regional extent (Thompson and Norton, 1968). It separates the higher grade metamorphic terranes of the Bronson Hill anticlinorium from the belt of higher grade domes of the Connecticut Valley-Gaspé synclinorium in eastern Vermont. The highest grade of regional metamorphism in the synclinorium in the Brattleboro area, staurolite-kyanite zone, is centered on the Guilford dome. Peak metamorphic conditions occurred late in the deformational cycle likely synchronous with and following the later stages of the doming deformation. During the earlier recumbent folding, metamorphic conditions likely did not exceed those of the garnet zone. A small diopside-bearing contact aureole is developed adjacent to the Black Mountain Granite (Stops 5 and 8).

## TIMING OF DEFORMATION AND METAMORPHISM

Deformation and metamorphism in the Connecticut Valley-Gaspé synclinorium in southeastern Vermont are assigned to the Acadian Orogeny, although the recent determination by Hayward and others (1988) of a Carboniferous crystallization age for the Black Mountain Granite may indicate the deformation and metamorphism lasted longer than previously thought. Clearly most of the regional deformation had occurred prior to the intrusion of this granite. However, the granite body and many of the dikes and sills of granite that extend into the country rock have a weak foliation produced by the alignment of fine-grained micas. This foliation has a steep to vertical dip and roughly parallels the strike of the schistosity in the surrounding rocks. Previously, Naylor (1971) interpreted the ages of coarse unaligned muscovite flakes from the Black Mountain (377 and 383 $\pm$ 7 Ma, early Middle Devonian) as likely establishing a minimum age for this intrusion.

The new age determination for the Black Mountain Granite brings into question the relation between the timing of its intrusion and the rise of the Guilford dome. If the granite is a post-Acadian intrusion and cross-cuts the dome and surrounding rocks, as seems to be the case at the present erosion surface, then the fine-grained mica alignment in the granite would be evidence of younger, perhaps Alleghanian deformation which previously has not been identified in the area. On the other hand, if the intrusion of the Black Mountain Granite played a role in the rise of the Guilford dome and of the surrounding staurolite-kyanite isogradic surfaces, as was hypothesized by Hepburn and others (1984), then deformation and metamorphism during the Acadian Orogeny extended over a much longer time in southeastern Vermont than previously believed.

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## ITINERARY

Maps: Excursion stops will be in the Brattleboro Vt.-N.H. 15 minute quadrange. Geologic maps of the Brattleboro quadrangle (Hepburn and others, 1984) and of the State of Vermont (Doll and others, 1961) are available from the Vermont State Library, Montpelier.

Assembly point is the parking lot of the Howard Johnson Restaurant at the junctions of Routes 5, 9, and Interstate 91 north of Brattleboro Vt. (just off Interstate 91, Exit #3). Cars may be left here for the day and car pooling is encouraged.

Mileage

0.0 Exit parking lot of the Howard Johnson Restaurant, turn right (north) onto Vt. Route 5 at the traffic lights.

0.7 Overpass over Interstate 91.

0.9 Brattleboro-Dummerston town line.

1.2 STOP 1. Park in rest and picnic area on the east side of Route 5.

PUTNEY VOLCANICS. The Putney Volcanics (Trask, 1980) in this area consist of fine-grained, poorly foliated, light greenish gray quartz-plagioclase-muscovite phyllite and laminated granulite (used texturally) with interbedded gray slate. The granulites and feldspathic phyllites weather buff to light brownish gray, characteristic of feldspar-rich rocks. Many of the foliation surfaces have a notably silky sheen. Small, brownish pits where carbonate has weathered out are common. A few lenses of quartz-pebble conglomerate assigned to the Littleton Formation may be seen along Route 5 south of the highway pull-off but are much better developed at Stop 2. The rocks have been metamorphosed to the chlorite zone at this locality.

Continue north on Route 5.

1.4 Outcrop of Putney Volcanics to the east.

1.5 Outcrop of Putney Volcanics to the west.

2.1 Slate quarry in the Littleton Formation to the east.

2.3 STOP 2. Park on left (west) side of the road in small highway pull-off.

CONGLOMERATIC MEMBER, LITTLETON FM. & PUTNEY VOLCANICS. First examine outcrops of gray slate in the Littleton Formation on the east side of Route 5. Then cross Route 5 and walk approximately 0.1 mile north through woods to an abandoned chicken-yard near houses. (The chicken-yard has become increasingly overgrown in recent years since the chickens flew the coop.) Here the conglomeratic member of the Littleton Formation is exposed along with the phyllites and feldspathic granulite interbeds of the Putney Volcanics, immediately west of the conglomerate. The conglomerate contains both quartzite and slate pebbles in a slate matrix. (As this is the best exposure and type locality for the conglomerate, NO HAMMERING--PLEASE!)

The contact of this conglomerate with the Putney Volcanics represents the division between the "Vermont" and "New Hampshire" sequences in this area and has often been referred to informally as the "Chicken Yard line," named for these exposures. Cross bedding in laminated feldspathic granulitic beds of the Putney very near its contact with the Littleton Fm. at this locality (see if you can find them) and elsewhere in the Brattleboro quadrangle, along with possible load casts and apparent channeling in the conglomeratic beds of the Littleton (see Hepburn and others, 1984) consistantly indicate that the stratigraphy gets younger to the east, i.e. that the Littleton overlies the Putney. This implies the package of rocks between the Shaw Mountain Fm. and Putney Volcanics, the "Vermont" sequence, is pre-Littleton in age, likely late Ordovician or Silurian to earliest Devonian. Hepburn and others (1984) suggest the contact between the Putney and Littleton Fms. as exposed here is possibly an unconformity. However, an alternative interpretation that this contact represents a major pre-metamorphic thrust (Whately Thrust) that carries an older Littleton Fm. westward over younger Putney and Gile Mountain Formations has been advanced by Robinson and others (in Press) (see also discussion in Hepburn and others, 1984, p.142).

A few small porphyroblasts of light pink garnet can be seen in this outcrop. However, because of the high MnO content of these garnets, to 15.9 wt. percent, these rocks have been mapped as belonging with sub-almandine zone assemblages.

West of the abandoned chicken-yard a sequence of phyllites and feldspathic granulites similar to those at Stop 1 is exposed on the side of the hill.

Return to cars. Continue north on Route 5.

2.4 Road junction with dirt road on right, continue north on Route 5.

2.6 Roger's Construction Co. yard on right (east), possible alternate parking for Stop 2.

2.9 Dutton Pines State Forest.

3.4 Road junction with road to East Dummerston; continue on Route 5. Outcrop of Putney Volcanics to the west.

3.8 Road junction. Turn left (west) on road to East Dummerston and Dummerston Center.

4.7 Road junction in East Dummerston; continue straight.

4.8 Junction with road on right; continue straight.

4.9 Outcrop of Waits River Formation.

5.9 Dummerston Center. Turn sharp left (south).

6.0 STOP 3. Park along side of road.

NORTHFIELD FORMATION. Walk west to outcrops of the Northfield Formation exposed near the hinge area of the recumbent anticline above the Prospect Hill recumbent fold (see fig.1). The Northfield here is a gray, well-foliated mica schist with conspicuous garnet porphyroblasts and fewer porphyroblasts of biotite and staurolite. A few thin interbedded quartzites are also present.

Turn around; return north to Dummerston Center.

6.1 Dummerston Center. Turn left (west) on the paved road past fire station.

6.5 STOP 4. Park in road pull-off on north side of the road just before curve.

HINGE OF PROSPECT HILL FOLD, WAITS RIVER FM. & STANDING POND VOLCANICS. The Standing Pond Volcanics outline the northeasterly plunging hinge of the Prospect Hill recumbent fold at this locality (fig. 1). A 1/2 mile traverse will be made around this hinge by following the geologic contact between amphiboles of the Standing Pond Volcanics and schists, calcareous schists, and impure marbles of the Waits River Formation. This traverse presents an excellent opportunity to view a well-exposed hinge of a major recumbent fold. The contact is sharp and is easy to follow. The traverse begins just east of the pull-off near a very small creek along the eastern contact of the Standing Pond Volcanics. Follow this contact to the north through the woods and around the northeasterly plunging hinge of the recumbent fold, which closes on the lower south-facing slopes of Prospect Hill. Continue along the contact back southward (now the western contact of the Standing Pond with the Waits River). The paved road is again encountered about 1/4 mile west of the starting point.

Particular note should be made of the minor folds during the traverse. The most common folds are the $F_2$ generation, those formed congruously with the recumbent folding. These plunge NE and show a reversal of drag sense around the recumbent fold hinge; "M" folds are common near the axial surface of the major fold. A few $F_1$ minor folds that pre-date the recumbent folding, have the principal schistosity parallel to their axial surfaces, and are refolded by the $F_2$ folds, are visible in outcrops near the road.

If time and weather permit we will climb Prospect Hill for an excellent view from the open summit (perhaps lunch). Please be particularly careful on this traverse with litter and the indiscriminate use of hammers. We are able to make this stop only with special permission.

Return to cars; continue west on paved road.

6.7 Outcrops of the Standing Pond Volcanics in the hinge of the Prospect Hill recumbent fold.

6.8 Contact of the Standing Pond Volcanics with theWaits River Fm.

6.9 Junction with dirt road to south; continue straight on paved road.

7.4 Outcrop of aplitic dike associated with the Black Mountain Granite.

7.8 Junction with road from right (north); continue straight.

8.5 Road junction; turn sharp left onto dirt road.

9.3 STOP 5. Park by abandoned quarry buildings and follow path east to the abandoned Presbury-Leland granite quarry.

BLACK MOUNTAIN GRANITE. The Black Mountain Granite is a late synorogenic to post-orogenic pluton correlated with the New Hampshire Plutonic Series (Billings, 1956). It is commonly a very light gray, massive, two mica granite-granodiorite with small garnets as an accessory phase. Corundum (1-2%) is present in the norm (Hepburn and others, 1984). Note the weak foliation produced by the alignment of the fine-grained micas. Recently, Hayward and others (1988) obtained a Carboniferous crystallization age of 326 $\pm$ 17 Ma (Rb-Sr whole rock) and an initial $^{87}Sr/^{86}Sr$ ratio of 0.7118 $\pm$ 0.0008 for the Black Mountain. Previously, a Devonian age had been assigned to this pluton by Naylor (1971).

West- to northwest-dipping sheeting is well exposed in the quarry walls. Note particularly how the thicknesses of the sheets increases with depth.

STOP 5A. Walk northwest from the quarry to the banks of the West River where the contact of the granite with the surrounding Waits River Formation is well exposed. Dikes and sills of granite and aplite are abundant within a few hundred feet of the contact and likely indicate stoping as the mechanism for emplacement of the pluton. The dikes cross-cut bedding and the principal schistosity.

Some have a weak foliation roughly parallel to the regional schistosity but they clearly post-date the major deformation. The country rocks near the granite have been altered by contact metamorphism, in addition to having been regionally metamorphosed to the staurolite-kyanite zone.

Return to cars; turn around and retrace route north the the main paved road.

10.1 Junction with paved road; continue straight (north).

10.2 STOP 6. Park near the entrance to the covered bridge on the east side of the West River.

WAITS RIVER FORMATION. Outcrops typical of the Waits River Formation in the center of the Guilford dome are seen along the east bank of the West River under the bridge. The rocks are interbedded impure marbles, calcareous mica schists, and mica schists. Most of the minor folds present here are assigned to the $F_2$ stage and developed congruously with the large-scale recumbent folding. They were refolded into their present attitude by the rise of the Guilford dome.

Return to cars; proceed west across the covered bridge to Vt. Route 30.

10.3 Turn right (north) on Route 30.

10.8 Maple Valley Ski Area on left.

11.5 Outcrops of Barnard Volcanics on left, we will return to these after turning around.

12.2 Junction with road to Williamsville, just before bridge over Rock River. Turn around with care and return south on Vt. Route 30. (Note: garnet mica-schists typical of the Northfield Fm. are exposed on the south bank of the Rock River below the bridge.)

12.9 STOP 7. Park with care on the right shoulder beside large road-cuts.

BARNARD VOLCANICS. The Middle Ordovician Barnard Volcanics are exposed here in the Fall Brook anticline, which forms the core of the proposed recumbent anticline above the Prospect Hill fold (fig. 1). The anticline is overturned, with both limbs and the axial surface dipping moderately to steeply northwest. At this stop, the rocks include amphibolites and felsic gneisses characteristic of the Barnard. Minor amounts of rusty-weathering schist similar to those of the Cram Hill Member are also present in this anticline, but have not been shown separately on figure 1.

Continue south on Route 30.

14.2 Covered bridge on left; continue straight on Route 30.

14.9 Note Black Mountain and the granite quarry to the east across the West River.

15.3-.6 Outcrops of the Waits River Formation.

15.8 Iron bridge to left; junction of road to the right; continue straight on Route 30. Outcrops of granite in the brook to the west.

17.2 STOP 8. Park at side of Route 30 by the large road-cut on the right (west).

CONTACT METAMORPHOSED WAITS RIVER FORMATION. The Waits River Formation in the outcrop here, near the contact of the Black Mountain Granite, has been affected by contact metamorphism in addition to staurolite-kyanite grade regional metamorphism. Calc-silicates, particularly actinolite and diopside, are well developed in the impure marble beds. Diopside has not been observed in the Waits River Formation in the Guilford done area outside of the contact aureole of the Black Mountain Granite.

Continue south on Route 30.

18.8 Small roadmetal quarry in the Waits River Formation.

19.0 Outcrop of Waits River Formation.

19.7 STOP 9. Park at the left in the pull-off beneath the I91 overpass.

GILE MOUNTAIN FORMATION, MARBLE MEMBER. Exposures under the overpass are fairly fresh outcrops of the marble member of the Gile Mountain Formation, metamorphosed to the biotite zone. Impure marble beds (already starting to obtain the distinctive punky-brown weathering rind) similar to those in the Waits River Formation are interbedded with micaceous quartzites and phyllites. The marble member of the Gile Mountain Formation is similar in many aspects to the Waits River Formation except that it contains a higher percentage of quartzitic beds and occurs east of the Standing Pond Volcanics.

END OF FIELD TRIP

Continue south 1.5 miles to Brattleboro for junctions with major highways. To return to Keene, turn left (north) in Brattleboro on Route 5 and follow about 2 miles to the junction of Route 9 at the Howard Johnson Restaurant. Follow Route 9 east to Keene.

# Trip C-1: Geology of Mount Monadnock

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## ITINERARY

Text precedes Trip B-2. Trip consists of a hike to the summit of Mt. Monadnock, weather permitting. In case of rain we will visit lower elevations. Assembly point is Keene State College Commons Parking Lot, 8:30 a.m. Participants should bring water and lunches. Topographic maps: Monadnock 15' quadrangle OR Monadnock Mountain 7.5' X 15' quadrangle.

Mileage

0.0 Take Wyman Way to Main Street.

0.2 Right (S) on Main Street.

0.5 Left (E) on Rte. 101.

1.8 Hills to N and S are Clough Quartzite on east edge of Keene dome.

3.9 Fitzwilliam Granite south side of road.

4.4 Marlboro village -- poor exposures of Rangeley Formation.

4.6 Right (S) on Rte. 124. Stay on Rte. 124 to Mt. Monadnock.

5.7 Abandoned granite quarry to left (Fitzwilliam Granite).

6.5-6.9 Rangeley outcrops in woods, both sides of road.

7.8 Spaulding Tonalite north of road. Plutonic rocks here to foot of Mt. Monadnock, intruding the west limb of the Monadnock syncline.

10.4 Perkins Pond. View of Mt. Monadnock. Pass Troy Road, right.

11.2 Spaulding Tonalite on right.

11.3-11.4 Littleton Fm. on left, intruded by granite.

11.9 Turn left into parking area for White Arrow trail to Mt. Monadnock. We will visit Silurian units on the east limb of Monadnock syncline (Fig. 1), and then spend the rest of the day in Devonian Littleton Fm.

STOP 1: Proceed on foot up old toll road (Half Way House Road) 650 feet to outcrop, right side of road: Francestown Fm., bedding oriented N14E, 44 NW. Proceed 115 feet up road to small bedrock exposure in left road bank (Francestown) and go west into woods perpendicular to strike to more Francestown outcrops on small ridge.

Proceed N70W to stonewall, north 60 feet along wall, cross at flag, and go another 100 feet north up slope to conspicuous outcrop west of wall: lower part of the Warner Fm. Thinly bedded calc-silicate granulite. Pink quartz-garnet-clinozoisite-diopside layers, green actinolite-quartz-anorthite-microcline layers, gray to white quartz-epidote-diopside layers. Bedding N38E, 40 NW.

Continue NW perpendicular to strike about 100 feet to an outcrop of Fitzwilliam Granite at E-W stonewall and Warner Fm. 20 feet beyond. Continue NW perpendicular to strike 75 feet to outcrop of upper part of Warner Fm. and 18-inch granite sill. Follow strike NE 50 feet to

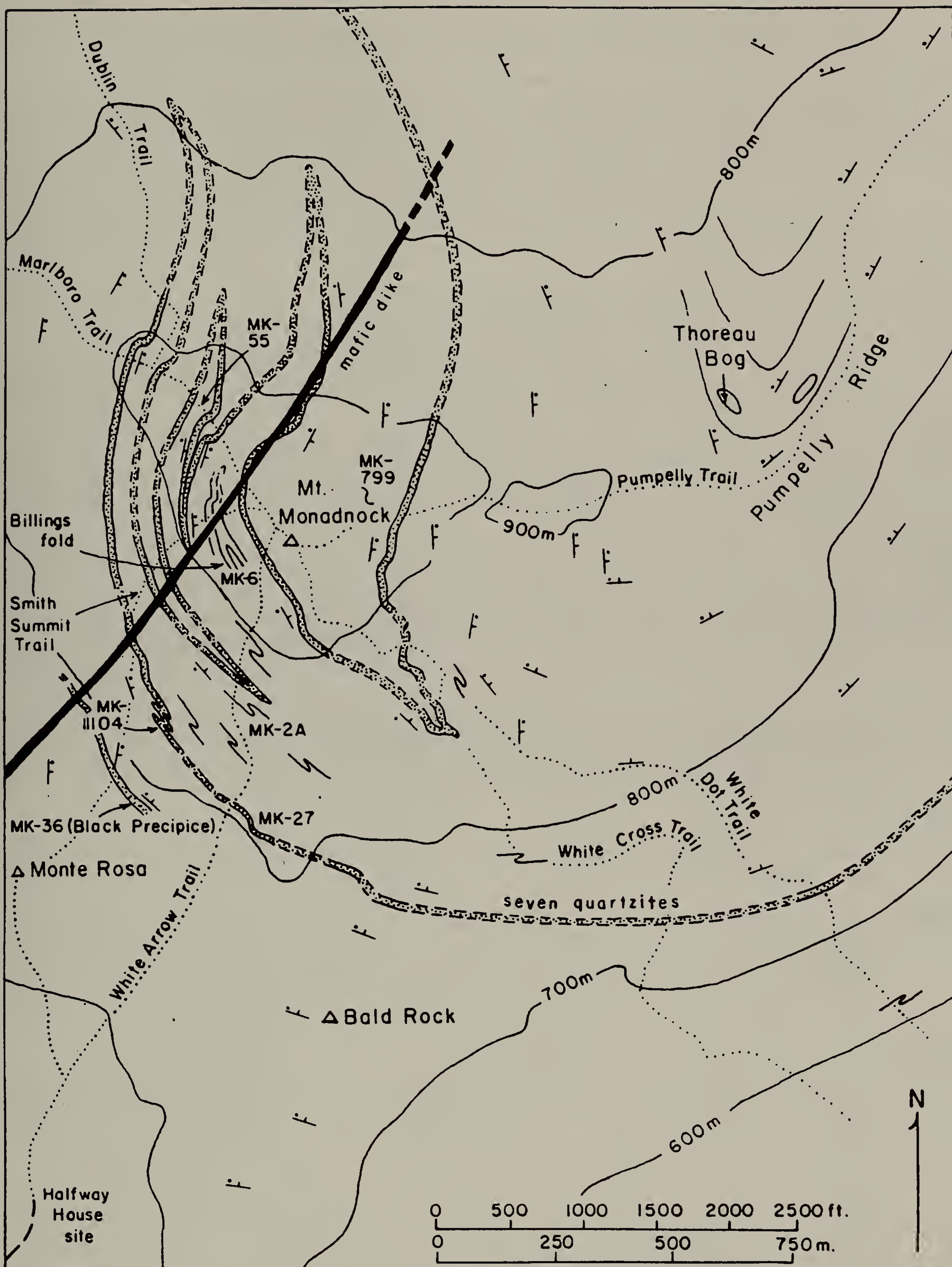

Trip C-1, Fig.1. Geology of Mt. Monadnock area. Gray pattern indicates outcrops. Stops 7 & 8 are between Stop 6 and summit (triangle). Formations abbreviated as on text Fig.3. Mmd—microdiorite dike; Mfg—Fitzwilliam Granite; Dst—Spaulding Tonalite; Dlu7q—Seven quartzites in the upper Littleton Fm.

more Warner, then perpendicular NW 60 feet to outcrop near ridgeline: upper Warner quartz-plagioclase-biotite granulite, with calc-silicate pod.

Follow contour of land east, passing round granite boulder, and at 220 feet another Warner outcrop (poison ivy abounds), and 200 more feet back to toll road. Continue up road.

Watch for Littleton Fm. outcrops at left edge of road. Just beyond here road levels out (elevation 516 m, 1695 ft).

| Elevation m | ft | |
|---|---|---|
| 546 | 1790 | At next steep pitch, ledge of atypical, somewhat rusty Littleton. Parker trail and spur to right lead 750 feet to excellent exposure of lower Littleton and view southeast (optional stop). |
| | | Beyond dip in road, Littleton outcrops and stonewall on right. |
| | | 600 feet farther on right, Littleton outcrops. |
| 576 | 1890 | 400 feet farther on right, glacially striated outcrop of Littleton. Striae and grooves oriented N55W. |
| 620 | 2035 | STOP 2. Freshly exposed outcrops in trail where trail leaves road: quartz-biotite-muscovite-garnet-sillimanite schist, typical of lower part of Littleton Fm. in Zone III. Quartzite layers are thin and widely spaced or absent. At clearing (old Halfway House site), schist is intruded by pegmatite. View of west side of Mt. Monadnock: bedding dips away from us; flat surfaces are joints. |
| | | Proceed through clearing on main trail. Where White Arrow trail starts up steeply over rocks, take Fairy Spring trail to left, across brook and past 1855 Fassett's Mountain House site. |
| 686 | 2250 | Trail levels out. Ledge to right with upright graded quartzite bed. |
| 707 | 2320 | Fairy Spring. Widely spaced, thin quartzite layers. |
| 719 | 2360 | Trail goes up over ledge with yellow blaze: upright graded bed and coticule layer. |
| 735 | 2410 | Bear right (up) on Monte Rosa trail. |
| 739 | 2425 | Bear left to Monte Rosa (white blazes). |
| 774 | 2540 | STOP 3: Monte Rosa. 180° view from Mt. Wachusett (S25E) to Mt. Ascutney (N25W). Low country to west is underlain by Fitzwilliam Granite and Spaulding Tonalite. Summit of Gap Mtn.(S30W) is made of Littleton and Silurian xenoliths (roof pendants?) surrounded by Spaulding. Troy (S65W) lies in the southern extension of the Monadnock syncline, which parallels the ridge of Little Monadnock to the left of Troy. Berkshire-Green Mountains on the western skyline: Greylock at S80W and Stratton Mountain at N70W.<br>On the highest part of Monte Rosa there are good examples of unaltered sillimanite pseudomorphs after andalusite, resistant to weathering. |
| | | In the next outcrop area toward Monadnock, beyond woods (near the "Tooth"): large isoclinal folds, tens of feet in amplitude, close both NW and SE. Foliation parallels the axial planes: N39W, 41 NE. |
| | | Continue on Smith Summit trail toward the top. Quartzite beds become |




Trip C-1, Fig.2. Geology of Mt. Monadnock summit, showing isoclinal folds marked by the "seven quartzites". Trip C-1 follows the Smith Summit Trail from Monte Rosa to the summit.

| m | ft | |
|---|---|---|
| | | more abundant, as well as pegmatite and aplite sills. Go past first trail to Black Precipice, which leads to the base of cliff. |
| 798 | 2620 | STOP 4: Take second spur trail to top of Black Precipice: seven closely bedded quartzite beds (N63W, 26 NE) of the upper part of the Littleton Fm. We will see these layers again several times as we cross isoclinal folds on our way to the summit (Fig. 2). Here the beds are right-side-up, although the seven quartzites themselves are usually not graded, and one must seek topping directions in the adjacent layers. |
| | | Return to Smith Summit trail. |
| 811 | 2660 | In the trail, seven quartzites in a large backfold: bedding N57E, 52 SE. |
| 835 | 2740 | Tourmaline vein. Tourmaline has replaced minerals in the schist, especially the andalumps. These veins are very common on the mountain, apparently coeval with the pegmatites. |
| 850 | 2790 | STOP 5: MK-1104. Follow ledges SE to seven quartzites in two tight isoclinal folds on the upright limb of a larger isocline. Beds are attenuated on limbs and thickened in hinges -- classic example of similar folds. |
| 855 | 2805 | Return to Smith Summit trail, where the seven quartzites cross the upper part of a large smooth outcrop area with well developed joint sets. Trail enters spruce woods; note a six-foot-wide gap in outcrop oriented N45E up and down slope. This marks the location of a microdiorite dike which crosses the mountain NW of the summit (see Stop 7). |
| | | In the next outcrop west of the dike, the trail crosses the seven quartzites, and continues across right-side-up graded beds. Watch for reversal in topping direction of graded beds as we cross a recumbent syncline which opens to the SE. |
| 887 | 2910 | Seven quartzites upside down. Trail follows them up to the north. Look for grading in nearby beds. Trail leaves the seven quartzites in a muddle of quartz veins and pegmatite. |
| 893 | 2930 | Next flattish area has good graded upright beds, coticule lenses, and curious structures which may be schist boudins between quartzite beds. Adjacent quartzite layers approach each other, often with a quartz knot in the boudin line. A boudin origin suggests some unusual ductility contrast conditions between the schist and quartzite. |
| 914 | 3000 | STOP 6: Fourteen quartzite beds at the Smith Summit trail, in the nose of an isoclinal fold. The two limbs can be seen going off across the mountain to the north: the upright limb to the NNW and the overturned limb to the NNE. If time and weather permit, we may follow this structure north to the Dublin trail, after visiting Stop 7. |
| | | STOP 7: Leave the trail, descending slightly toward the south, 150 feet to the bottom of a cliff with a sharp cleft. The microdiorite dike we crossed earlier is exposed in this cleft. It is about six feet thick, nearly vertical, and cuts across all folds in the Littleton. It is cut by tourmaline and pegmatite veins. It is parallel to granite dikes where it is last seen SW of Monte Rosa, and contact relations suggest it is about the same age as the Fitzwilliam Granite. |

| m | ft | |
|---|---|---|
| | | The dike is a fine-grained, weakly foliated, plagioclase-biotite-hornblende-quartz-ilmenite rock, with clots of biotite that give the weathered rock a spotted appearance. Some samples contain small corroded garnets (xenocrysts?). The plagioclase is $An_{45}$. Ilmenite is commonly rimmed by sphene and hornblende is rimmed by biotite. The dike is clearly post-tectonic and therefore younger than Acadian, yet it is somewhat metamorphosed. Could this be a result of the "Permian disturbance" of southern New England? |
| | | Return to the Smith Summit trail at Stop 6, and continue up trail. |
| 945 | 3100 | STOP 8: "Billings' fold", on cliff to right, was pictured in first edition (1942) of Marland Billings' Structural Geology. This fold lies structurally within the uppermost of three southeast-opening isoclinal synclines defined by the seven quartzites (Fig. 2). Fold axis: N58E at 32°. Foliation and axial plane: N16W, 36 NE. |
| 963 | 3160 | Continue to summit of Mt. Monadnock. Views of Dublin Pond (Warner Fm.) and Skatutakee Mtn. (Kinsman Granite in the Cardigan pluton) to the north. To the northeast, Pumpelly Ridge extends as the east limb of the Thoreau Bog syncline. At the summit, bedding dips NE. On Pumpelly Ridge, bedding dips NW. Crotched Mtn. and the Pack Monadnock Range can be seen in the middle ground to the east. Littleton Fm. is exposed on North Pack (Duke, 1984) in what may be the next higher synclinal nappe above the Monadnock syncline. Thorndike Pond can be seen closer to the mountain in the east; layers of Kinsman extend from the Cardigan pluton south past Thorndike Pond toward the Coys Hill Granite in Massachusetts. |
| | | Return to the vehicles via the White Arrow and Sidefoot trails and the old toll road. If time permits, a loop across Bald Rock and past the old graphite mine (Cliff Walk trail) is worth the effort. Return to the Halfway House site via Do Drop or Hedgehog trails. |

# THE SKITCHEWAUG NAPPE IN THE MASCOMA AREA, WEST-CENTRAL NEW HAMPSHIRE

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## INTRODUCTION

The highway cuts on Interstate 89 provide an excellent opportunity to see some of the more interesting features of the geology of the Bronson Hill anticlinorium in the area between the Mascoma and Croydon domes. The highway crosses at the structural depression between the domes. This depression displays a segment of the Skitchewaug nappe, and a floored outlier of the Bethlehem Gneis of the Mount Clough Pluton.

The area has been of interest since the early work of C. H. Hitchcock (1877, 1890, 1908, 1912). Later work by C. A. Chapman (1939, 1942, 1952); by F. C. Kruger and Daniel Linehan (1941); and by J. B. Lyons (1955) laid the ground work for the present interpretation (Thompson, 1954, 1956; Thompson *et al.*,, 1968; Robinson *et al.*, 1979). The discovery of fossils at Skitchewaug Mountain, Vermont (Boucot *et al.*, 1958; Boucot and Thompson, 1958, 1963) and later in the vicinity of the Croydon and Mascoma domes provided some of the key information leading to the identification of the Skitchewaug nappe as a major structural feature of the Bronson Hill anticlinorium. The emplacement of this nappe represents an early stage of the tectonic history of the region, antedating the rise of the domes.

The stratigraphic sequence shown in figure 1 is essentially that worked out by Marland P. Billings (1937) in the Littleton-Moosilauke area, farther north, but on-strike with figure 1. An early Devonian age for the Littleton Formation is established on the basis of abundant fossils near Littleton, NH (Boucot and Arndt, 1960; Boucot and Rumble, 1980). The Silurian age of the Fitch Formation is based on fossils at the Fitch farm and vicinity, also near Littleton (Harris *et al.*, 1983). The dating of the Clough Formation as early Silurian, however, is based mainly on fossils immediately south of figure 1, at Hetty Brook and Beaver Brook in the Croydon Mountain northwest of Newport, NH (Boucot and Thompson, 1963). These localities, unfortunately, are inaccessible to an NEIGC group, and we must therefore be content with the less-well-preserved material in the area of figure 1.

The only rocks to be seen on this trip that are older than the fossiliferous Clough, are assigned to the Partridge (Quimby?) Formation. It is of probable Ordovician age, but may include some younger strata. As pointed out by Hepburn *et al.* (1984, p. 143, see also Trip A-2 this volume) the pre-Littleton rocks in the Skitchewaug nappe represent a more southeasterly lithofacies than their correlatives mantling the domes. The rocks in the nappe stratigraphically beneath the fossiliferous Clough quartzites, include a variety of pelites and conglomerates shown on earlier maps as part of the Clough. These (and also some of the pelites previously mapped as Partridge or Littleton) are now regarded as correlative with parts of the Rangeley and Perry Mountain Formations of Hatch *et al.* (1983).

In the opinion of the writer the fossiliferous quartzites at the top of the Clough Formation correlate with those of the uppermost Perry Mountain Formation of Hatch *et al.* (1983). The Fitch Formation, commonly containing rusty-weathering schists and calc-silicates in its lower part, and non-rusty calc-silicates in its upper part, would then be correlative with the Smalls Falls and Madrid Formations, respectively. Both Hatch *et al.* and the writer agree that the lower units of their Silurian section are cut out by onlap northwestward toward the Bronson Hill anticlinorium, a positive area during Silurian sedimentation. The writer respectfully suggests, therefore, that the Blanchard Pond fossil locality, that has been used to date the Rangeley Formation, may in fact belong to the Perry Mountain rather than the Rangeley, and that the Rangeley is simply missing at Blanchard Pond. The faunas at Blanchard Pond are essentially the same as those at the Croydon Mountain, and the Skitchewaug Mountain localities farther south. The lower parts of the Clough and Perry Mountain Formations and all of the Rangeley, could thus be Llandovery or <u>older</u>, possibly including some Ordovician strata.

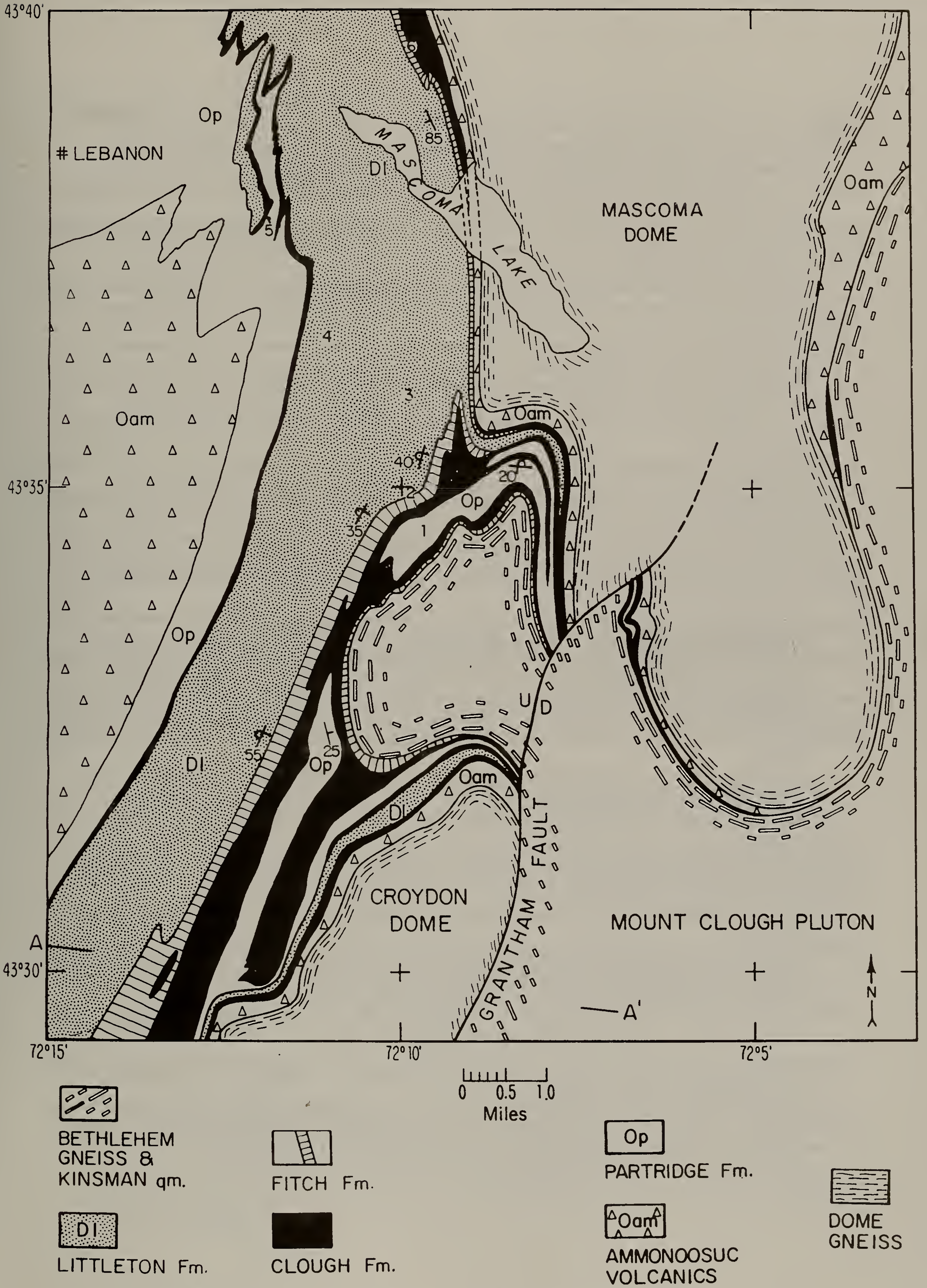


Figure 1. Geologic map of the Mascoma Lake area, New Hampshire. Modified from Thompson *et al.* (1968) and Chapman (1939, 1942, 1952). Numbers show stop localities. Line A-A' is section line of Thompson *et al.* (1968, Pl. 15-1b).

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## ITINERARY

Place of assembly: Exit 15 (Montcalm) on I-89 between Lebanon and Grantham, New Hampshire. If northbound on I-89 turn left at foot of exit ramp, pass beneath highway, then turn left again at T-intersection beyond southbound ramps. Proceed south 1.3 mi parallel and adjacent to I-89 to turnaround and parking area. Excess vehicles will be stored here. If southbound on I-89 turn right at foot of exit ramp, then left immediately at T-intersection. Proceed 1.3 mi south as above. Park and set trip mileage at zero.

STOP 1 - Outcrop 1. Quartzite of the Clough including the fossiliferous upper calcareous zone in the hinge region of the Skitchewaug nappe (Boucot and Thompson, 1963, Locality 9, on strike with Outcrop 6, and Locality 7, with Outcrops 1 and 2). Most of this outcrop consists of massive quartzite and quartz-pebble conglomerate separated by thin highly foliated layers of muscovite-garnet schist. In some places the quartzite beds appear to be boudined and the schist injected between quartzite masses. Overall the bedding dips gently to moderately northwest, whereas the dominant schistosity in the schist interbeds dips gently to steeply southeast, apparently parallel to axial planes of early (nappe stage) folds. Two tabular masses of biotite amphibolite dip moderately to steeply southeast. They are roughly parallel to the axial plane foliation in the schist and appear to truncate bedding, so they are probably dikes. A strong mica lineation on schist foliation planes and a strong hornblende lineation on amphibolite foliation planes trends N30-50W in reasonable conformity with features assigned to the dome stage in adjacent outcrops.

At the northern end of Outcrop 1, the quartzite is overlain by a well-bedded zone of diopside- and grossular-bearing calc-silicate rocks, fine-grained quartzites, and sulfidic mica schists, assigned to the upper member of the Clough Formation. Some calc-silicate rocks contain obvious shelly fossils, commonly replaced by diopside or grossular. DO NOT HAMMER: Plenty of loose material at Outcrop 2. In this vicinity there are several subhorizontal early folds in bedding that trend N20-40E with southeast dipping axial surfaces. As will be seen at Outcrop 2, these are equated with the hinge direction of the Skitchewaug nappe. Also note some late chevron folds in mica schist with steeply dipping axial surfaces.

Climb onto top of north end of Outcrop 1, observing weathered fossils en route, then traverse grass and dirt slope south to low point in fence, and climb into quarry which lies above and south-southeast.

STOP 1 - Outcrop 2: Rocks exposed in the quarry include typical quartzite and mica schist of the Clough Formation, the fossiliferous, bedded quartzite and calc-silicate member of the Clough, dikes of amphibolite, and a small post-metamorphic diabase intrusion

As you face upward into the quarry you are looking west-northwest up the southeast-dipping axial surfaces of a series of minor recumbent folds related to the frontal region of the Skitchewaug nappe. The fold axes are subhorizontal and trend N20-45E approximately at right angles to your line of view. At the mouth of the quarry you are standing in the quartzite member, which is also exposed on the left- and right-hand sides. The entire top and back of the quarry is composed of the fossiliferous bedded quartzite and calc-silicate member that lies in a more frontal position in the nappe. Unfortunately, most contacts between the lower quartzite and the upper quartzite and calc-silicate members are either covered by debris or exposed on badly altered surfaces. The amphibolite is clearly a dike parallel to the axial surfaces of the northeast-trending nappe stage folds. It is possible that the dikes post-date the folds, but we prefer to think that the fold axial surfaces where localized by the dikes, because such dikes, though present in the Fitch, are unknown in the overlying Littleton Formation, and are presumed to be of Silurian age. A moderate to strong northwest-trending mineral lineation, equated with the dome stage in adjacent outcrops, is evident throughout the quarry.

STOP 1 - Outcrop 3 - 8: The next six outcrops involve a hike of about 5.2 km over rough terrain with a vertical relief of 200 m. The feasibility of this hike is contingent on weather conditions. If weather does not permit, return to the bicycle path and proceed south along it about 0.8 km where you can pass through a hole in the fence to the prominent quartzite ledges of Outcrop 9. If weather does permit, on the other hand, ascend ridge north of quarry (Outcrop 2) in a southwesterly direction to the 578 m summit. The route passes over calc-silicate rocks and associated schists and granultes of the Fitch Formation, on into the fossil zone and the massive quartzites of the Clough, and thence into the rusty schists, amphibolite, and minor quartzites of the Partridge Formation. The summit (Outcrop 3), is amphibolite, possibly intrusive into the Partridge Formation.

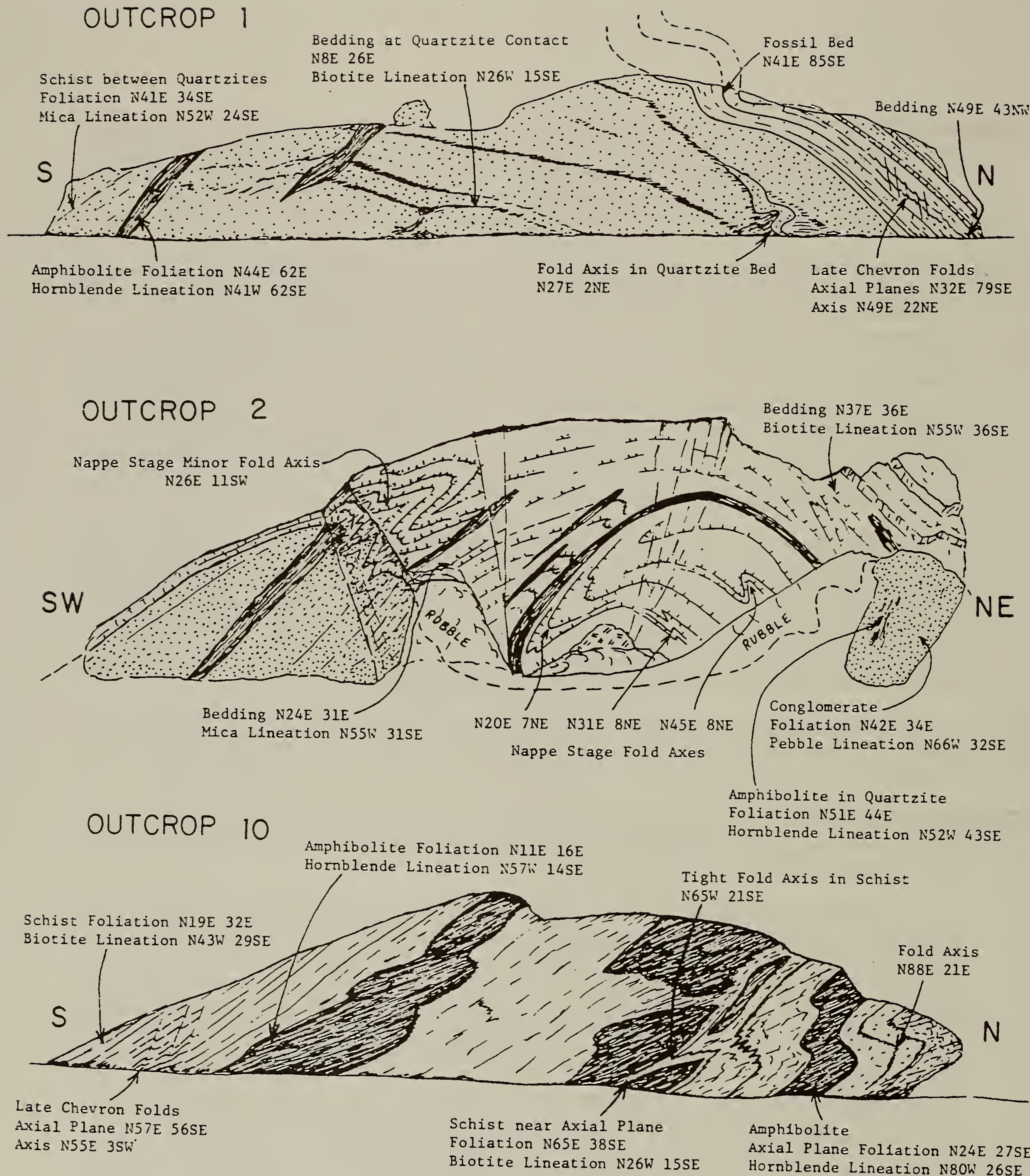


Figure 2. Field sketches (by Peter Robinson, 1979) of outcrops at Stop 1.

Outcrop 1. Clough Quartzite including upper, fossiliferous, quartzite and calc-silicate member in frontal region of Skitchewaug nappe, cut by dikes of amphibolite. Dominant folds and axial plane foliation seem to be of the nappe stage, but there are also abundant dome stage mineral lineations. Length of outcrop approximately 45 meters.

Outcrop 2. View into quarry in Clough Quartzite, frontal region of Skitchewaug nappe. View is looking WNW approximately parallel to the transport direction of the nappe. Quartzite member in foreground to right and left (stippled) is succeeded, up along the axial surfaces, by the fossiliferous quartzite and calc-silicate member. Both members are cut by amphibolite dikes approximately parallel to nappe stage axial surfaces. There is a small plug of post-metamorphic diabase at base of wall in center of quarry. Width of view in foreground is about 45 meters. Back wall of quarry is about 30 meters.

Outcrop 3. Partridge Formation in core of Skitchewaug nappe with abundant dome stage folds. Length of outcrop approximately 30 meters.

Proceed SW from the 578 m summit across lower ground, crossing the Clough, and ascending to the 589 m summit in calc-silicate rocks (note concretions) and associated rocks of the Fitch Formation. (Outcrop 4). The Fitch Formation also contains amphibolites, possibly intrusive, to the west and south, along crest of ridge. From the 589 m summit proceed SSW along high ground, along or near the Clough-Fitch contact, to NW base of a 590 m knob, and ascend SE to its summit in quartzites of the Clough (Outcrop 5). The quartzite here has pelitic partings containing almandine garnet to 1 cm. From here we proceed SSW, along or just W of ridge crest, following Clough-Fitch contact in search of fossils. At about 2.7 km from the start of hike is old town road crossing ridge. (This is the old road from East Plainfield to North Grantham). Knoll just north of road at crest (Outcrop 6) is fossil locality 7 of Boucot and Thompson (1963). Much of the best material, however, was removed by them. The terrain, then freshly logged, is now densely overgrown. From here we follow the old road E over quartzite with minor schistose partings. At about 0.3 km E of crest, drop south to E-flowing gully. Outcrop 7 (locality 6 of Boucot and Thompson) is on north slope of gully. The Clough-Fitch contact here tops east. We will then visit exposures of Bethlehem gneisses on south slope of 513 m hill NW of Leavitt Pond, and then cut N around W ridge of hill to road,and follow it E to N end of Leavitt Pond. LUNCH!

After lunch we will proceed NNE about 0.4 km in search of Outcrop 8, a fossil locality at Clough-Fitch contact (tops again east), near western margin of Bethlehem Gneiss. From here the route continues north and northeast over scattered quartzite outcrops of the Clough Formation 1.3 km to bicycle path.

STOP 1 - Outcrop 9: Pass through hole in fence and climb to top of outcrop away from highway face. Quartzite of the Clough Formation on upper (right-side-up) limb of Skitchewaug nappe. Massive to weakly bedded quartzite, muscovite schist, and minor conglomerate. Bedding strikes N30-40E and dips 30-45 SE. Mica lineation and apparent pebble elongation lineation trends N60-70W.

Large exposures to east across I-89 and on the mountain above are Bethlehem Gneiss of the Mount Clough pluton that lies structurally above. Minor amounts of sillimanite-mica schist of the Littleton Formation, granulites, and calc-silicate rocks of the Fitch Formation, and quartzites occur of the Clough below overhangs along the west face of the mountain. The dominant mineral lineation in all these rocks has the same N60W trend and is believed to have formed during the dome stage of deformation in the compressed zone between the Mascoma and Croydon gneiss domes.

Look northwest down I-89. First two outcrops are Partridge Formation in core of Skitchewaug nappe, the third outcrop is quartzite of the Clough (Outcrop 1) in the anticlinal hinge region. Pass back through hole in fence onto bicycle path. Proceed northwest on path to Outcrop 10.

STOP 1 - Outcrop 10. Partridge Formation in core of Skitchewaug nappe. Rusty-weathering pyrrhotite-mica-garnet-fibrolite schist with subordinate biotite, amphibolite, and fine-grained felsic gneiss. All large folds with moderately inclined axial surfaces in this outcrop, as well as in the large outcrop across I-89 to the east, appear to trend N50-80W parallel to the dominant mineral lineation and are believed to have formed during the dome stage of deformation. At the extreme south end of the outcrop is an example of late chevron-type folds with steeply inclined axial surfaces.

Proceed northwest along bicycle path past Outcrop 1 to cars.

STOP 2. Littleton Formation. Staurolite-sillimanite mica schist. Bedding obscure. Foliation strikes N40E, dips 40SE. Dome stage mineral lineation trends N40W. Fibrolitic sillimanite is best seen with hand lens on upper surface of outcrop.

MIleage

| | |
|---|---|
| 0.0 | Reboard vehicles. Proceed northwest |
| 1.0 | Littleton Formation, left |
| 1.1 | Montcalm underpass. Take northbound ramp onto I-89. |

1.3 Montcalm entrance ramp, northbound. Stop right.

STOP 3. Littleton Formation (staurolite zone) with coarse staurolite. Mesozoic lamprophyre dike. (Brief stop - 10 minutes). Proceed north on I-89.

1.7 Littleton Formation, left

2.0 Littleton Formation, right. Garnet zone

2.5 Littleton Formation, right. Slaty phyllite, biotite zone

2.7 Take Exit 16 - Purmont. Cross trace of Clough Quartzite (Hardy Hill), upper limb of Cornish nappe.

2.9 Turn right, then right again to end (3.1 mi) of road parallel to I-89. Walk to outcrop SE.

STOP 4. Phyllite of Littleton Formation. Metamorphic grade has fallen off markedly since Stop 3. There are rare, tiny garnets and no staurolite. Return to cars and drive to northbound ramp (3.3 mi) and onto I-89.

3.5 Lebanon City Line. Roadcuts over next 1.5 mi are in rusty, carbonaceous phyllites of Partridge Formation. Rust is from weathering of pyrrhotite.

5.2 Take Exit 17, turn right (East) on US-4.

5.4 Turn right on Stony Brook Road.

6,1 Large outcrops right and smaller one left. Turn cars and park.

STOP 5. The more easterly cuts are in rust schists and horblendic rocks of the Orfordville Formation (Chapman, 1939; Hadley, 1942), here regarded as correlative with the Partridge. The rocks immediately west are part of the Hardy Hill Quartzite of Chapman (1939), now regarded as possibly correlative with the Clough. The Hardy Hill includes quartzites, phyllites, and quartz-pebble conglomerates, some with a phyllitic matrix. Cross-bedding tops west toward phyllites of the Littleton Formation exposed in the more westerly cuts. Return by cars to US-4.

6.8 Turn right on US-4.

8.3 Cross Mascoma River.

9.3 Turn left (north) on Ruddsboro Road.

10.2 Turnout just past bridge, park cars.

STOP 6. Follow Lebanon-Hanover townlne ESE about 0.2 km to clearings. ENE 0.3 km across clearings, are exposures of gray schists of the basal Littleton. Immediately E in old sugarbush are calc-silicates and associated rocks of the Fitch Formation. The fossil zone in the upper part of the Clough Formation is exposed farther east, across the old town road in the woods at ±400 m elevation. Pelmatozoan columnals were found here by the writer and G.J. Wasserburg in 1962. Later material collected by A. J. Boucot in 1966 yielded a highly deformed cup coral and a possible *Stricklandia*. The fossil zone can be traced farther north into and Hanover and south, across the northeast corner of Lebanon, into Enfield.

If time permits, we can ascend to the nearby open ledges of Clough quartzites with excellent views. Bears have been haunting here this year.

END OF TRIP

# STRUCTURE AND METAMORPHISM FROM JAMAICA TO THE ATHENS DOME, VERMONT

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## INTRODUCTION

Western New England was visited by two major Paleozoic deformations, the Ordovician Taconic orogeny and the Devonian Acadian orogeny. The effects of the Taconic orogeny are most clearly recognized in an elongate belt along the border of New York with Connecticut, Massachusetts, and Vermont and into north-central Vermont (as shown on Fig. 2, compiled by Laird,1988). Taconian deformation is characterized by westward directed thrusting (e.g. Zen 1967; Stanley and Ratcliffe, 1985). In contrast, Acadian deformation and metamorphism are most intense in New Hampshire and central Massachusetts and Connecticut, and crustal shortening is usually attributed to large-scale recumbent folding (e.g. Thompson et al., 1968; Robinson and Hall, 1980).

There is a zone of overlap in north-central Vermont, southeastern Vermont, and western Massachusetts and Connecticut where both the Taconic and Acadian orogenies have left their mark. As many geologists working in this zone have discovered, it is not always easy to determine which orogeny is responsible for specific structural and metamorphic features. Yet, if we are to develop reasonable tectonic reconstructions of western New England, we must sort out the physical conditions of each orogeny. The purpose of this field guide is to describe the evidence in southeastern Vermont for polydeformation and polymetamorphism, emphasize the strengths and weaknesses of the data, and make some suggestions on how to distinguish between the effects of the Taconic and Acadian orogenies.

An important aspect of the field trip is to assess (and debate?) stratigraphic and contact relationships. The field trip crosses units mapped by Doll et al. (1961) as the Mount Holly gneiss, the Bull Hill gneiss, and the Tyson, Hoosac, Pinney Hollow, Ottauquechee, Stowe, and Missisquoi Formations (see Table 1, Trip B-6 herein). Our trip complements that of Rosenfeld et al. (1988), and our last two stops are the same as Stops 3 and 6 of their trip.

This year's gathering of the NEIGC marks the twentieth anniversary of the publication of Studies of Appalachian Geology: Northern and Maritime (Zen et al., eds., 1968), more popularly known as the "Billings Volume". Twenty years after its appearance, this landmark collection of papers still remains the natural starting point for geologists interested in the northern Appalachians. The issues we wish to discuss were clearly identified in several of the articles contained in that volume (e.g. Albee, 1968; Rosenfeld, 1968; and Thompson and Norton, 1968). A great deal of new information is available to help address these issues (or at least to fuel debate), which we plan to summarize, but those seeking definitive answers at this time will be disappointed. Our regret at not being able to provide more complete structural and metamorphic histories of the region is tempered by our awareness that better geologists than ourselves have been at work on these problems for some time, and they still have many questions. Our goal is to stimulate not to satisfy.

## DEFORMATION

Rocks in southeastern Vermont are on the east flank of a major anticlinorial structure, the Green Mountain massif (Fig. 1). Lithologic contacts and deformational fabrics dip moderately to steeply to the east, except around the Chester and Athens domes where they dip gently to moderately away from the cores of the domes. Map-scale structures clearly reveal a history of multiple deformations: an early stage of recumbent folding and a later stage of doming (Doll et al., 1961; Rosenfeld, 1968). Because these structures involve Silurian and Devonian rocks, these deformations have been attributed to the Acadian orogeny.

At the outcrop scale the deformation fabrics clearly record multiple episodes of synmetamorphic deformation. An early schistosity (maybe not the first deformation fabric) is characteristically overprinted by a crenulation cleavage, and this younger fabric is often cross-cut by a spaced cleavage with only limited recrystallization parallel to it.

Acadian-age structures have dominated the attention of geologists in southeastern Vermont. Acadian deformation took place at garnet to kyanite grade in this region (Thompson and Norton, 1968) and produced large-

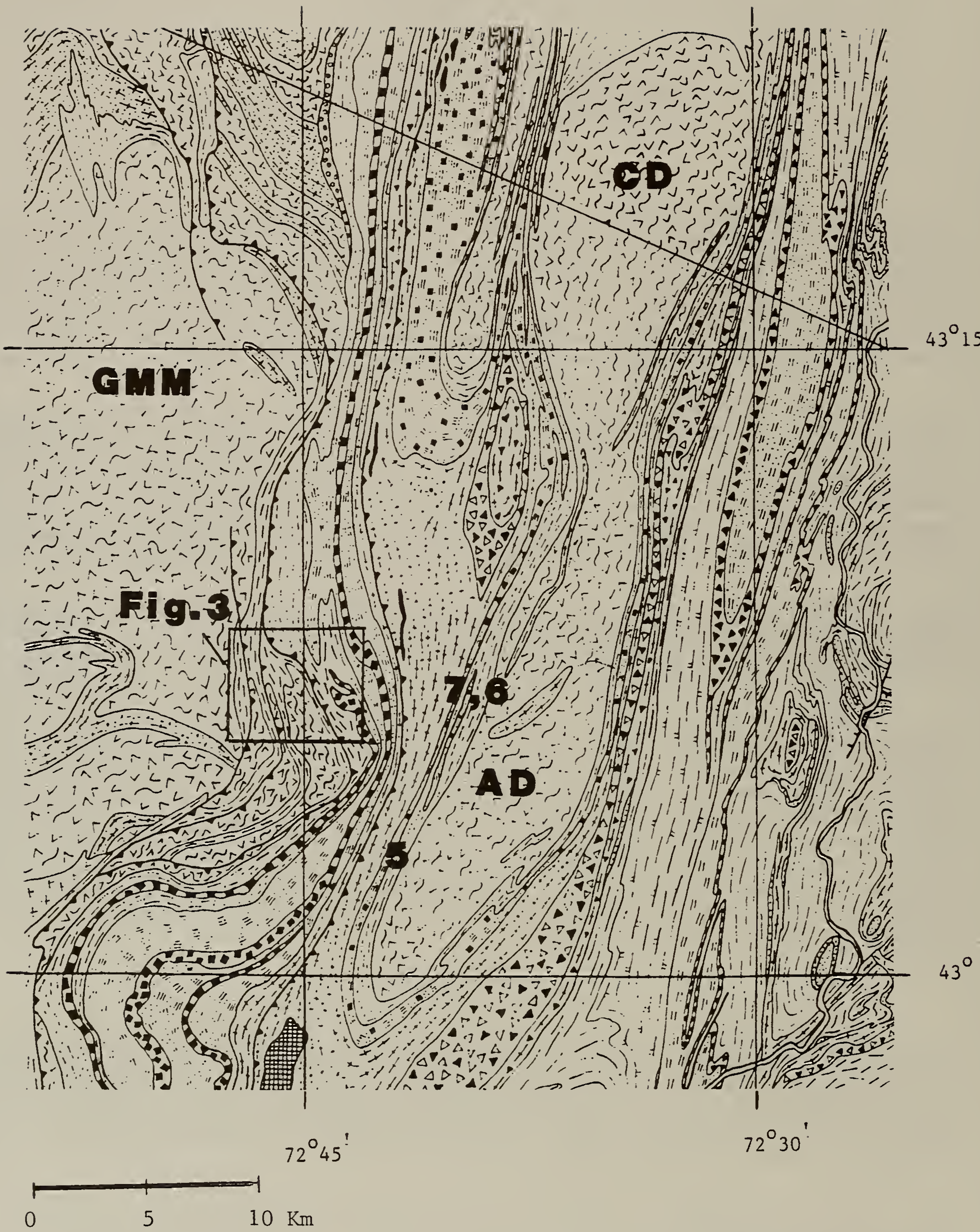


Figure 1: Geologic map of southeastern Vermont between the Green Mountain massif (GMM) and the Athens (AD) and Chester (CD) domes, with permission from Thompson et al. (1986). The same map and units as shown by Rosenfeld et al. (this volume, Fig. 2). Location of Figure 3 (Stops 1 - 4) is indicated.

scale ductile structures. Overprinting of older structures (assuming they really exist) was thorough. It seems that the first suggestions of Taconic deformation east of the Green Mountain massif stemmed from the quest for the "Taconic root zone", or a palinspastic source for the rocks now found in the Taconic klippen and emplaced during the Middle Ordovician onto coeval rocks of the Cambrian to Ordovician carbonate platform. Prindle and Knopf (1932) and Skehan (1961, 1972) mapped faults on the southeast margin of the Green Mountain massif which were candidates for this root zone, but the field evidence for Taconic deformation in southeastern Vermont was not compelling.

Rosenfeld (1968) described garnets from the Cambrian Pinney Hollow Formation on the west side of the Athens dome that contained unconformity textures unlike anything he saw in younger Silurian and Devonian rocks. (We use the age assignments of Doll et al., 1961 but are aware that fossil control is poor across our route and that paleontological studies in the Connecticut Valley trough are ongoing.) He made a bold suggestion that the early stage of garnet growth dated from the Taconic orogeny. This seems to be the first solid piece of evidence that the effects of pre-Acadian, possibly Taconic, orogeny extended to southeastern Vermont.

Karabinos (1984a) also found garnets with unconformity textures in high-alumina schists in the Hoosac Formation near Jamaica, Vermont, on the east flank of the Green Mountain massif (Fig. 2). Karabinos (1984b) used thin section textures and garnet zoning to show that the two stages of garnet growth were separated by a retrogression which partially resorbed first-stage garnet. He mapped thrust faults in the Jamaica area (Fig. 3) and argued (and continues to tell anyone who will listen) that thrusting of hot rocks to a structurally higher and cooler environment cut short the first prograde metamorphism. Thus, if the early stage of garnet growth really is Taconic and if thrusting was coeval with it, the thrusting is Taconic. Thermal modelling (Karabinos and Ketcham, 1988) suggests that such temperature fluctuations during thrusting in metamorphic belts are possible.

Mapping in the Berkshire massif and east of it during the late 1960's and 1970's by N.M. Ratcliffe, D.S. Harwood, R.S. Stanley, S.A. Norton, and N.L. Hatch was a major turning point in our understanding of the geology of western New England (see Stanley and Ratcliffe, 1985, for references). These geologists showed that the older Cambrian and Ordovician rocks contained structures not found in the Connecticut Valley trough and demonstrated that thrust faults were pervasive in western Massachusetts. These structures are presumably of Taconic age.

Since the late 1970's R.S. Stanley and his students at the University of Vermont have been mapping in pre-Silurian rocks in central and northern Vermont, and they also recognize numerous thrust faults (again, see Stanley and Ratcliffe, 1985, for references). Thrust faults have also been recognized in southeastern Vermont (e.g. Zen et al., 1983; Karabinos, 1984a; Thompson and McLelland, in press). The synmetamorphic thrust faults mapped in western Massachusetts and in Vermont are generally attributed to the Taconic orogeny (e.g. Stanley and Ratcliffe, 1985) although Ratcliffe (1979) has suggested an Acadian age for some thrusts. In some cases the evidence for a Taconic age for thrusting is good (i.e. Ratcliffe and Hatch, 1979; Sutter et al., 1985) but in many cases the evidence does not exclude an Acadian age. The common tendency to attribute thrusts throughout western New England to the Taconic orogeny appears to originate from an ingrained bias that the Taconic orogeny was dominated by thrusting and that the Acadian orogeny was dominated by recumbent folding. The next generation of research on thrusting in western New England must include tools to date fault movement.

## METAMORPHISM

The effects of Acadian metamorphism are obvious in southeastern Vermont where Late Precambrian to Devonian cover rocks contain magnificent porphyroblasts of garnet, staurolite, kyanite, and amphiboles. Harper (1968) used the K-Ar method to show that "the" metamorphism was Devonian in age and also demonstrated that Ordovician metamorphism occurred in the Taconic region, but it was unclear how far east Taconic metamorphism extended. As noted in the previous section, Rosenfeld (1968) used garnet inclusion textures to argue that Taconic metamorphism reached as far as southeastern Vermont. Also, Albee (1968) recognized that a younger metamorphism had been superimposed on an older metamorphism in northern Vermont, and Lanphere and Albee (1974) used 40Ar/39Ar ages to verify this assertion and to demonstrate that the early metamorphism was Taconic and the later metamorphism was Acadian. Two other studies using 40Ar/39Ar ages (Laird et al., 1984; Sutter et al., 1985) have convincingly shown that both Taconic and Acadian metamorphism occurred in Vermont and western Massachusetts, but so far only Devonian ages have been found in southeastern Vermont above the basement.

The lack of Ordovician ages from southeastern Vermont may reflect either thorough Acadian thermal overprinting or a lack of Ordovician metamorphism. We favor the former interpretation in accord with Rosenfeld

(1968) and Rosenfeld et al. (1988). Evidence for polymetamorphism in southeastern Vermont from both pelitic and mafic schists is now widespread and can be most easily interpreted as the result of separate periods of heating.

Pelitic schists: Rosenfeld's (1968) unconformity textures in garnets from the Pinney Hollow Formation on the west side of the Athens dome (STOP 6, Rosenfeld et al., 1988) demand two separate periods of garnet growth, but it is possible that both stages of garnet growth are Acadian. Cheney (1980) also presented evidence for polymetamorphism in high-alumina schists from western Massachusetts along strike to the south. Karabinos (1984b) described unconformity textures from high-alumina schists in the Hoosac Formation near Jamaica, Vermont (Stops 2 and 4), and zoning anomalies strongly suggest that the two stages of garnet growth were separated by a period of partial resorption of first-stage garnet. Downie (1982) reported garnets with unconformity textures in the Chester dome, and Hawkins and Skehan (1985) also found evidence for two stages of garnet growth near the southeastern margin of the Green Mountain massif. In a senior thesis project at Williams College, Cook (1988) sampled high-alumina schists from around the Chester, Athens, and Wilmington domes and found that the unconformity texture is widespread in southeastern Vermont.

Cook and Karabinos (1988) created two isograd maps following a method suggested by Thompson et al. (1977). The first isograd map is based on mineral inclusions in first-stage garnet, and the second is based on mineral inclusions in second-stage garnet and the matrix assemblage. The second isograd map appears to reflect peak Acadian metamorphism. The first isograd map may record either Taconic metamorphism or an early stage of Acadian heating. Clearly what is wanted is some method to date both stages of garnet growth.

Rosenfeld is once again in the vanguard and involved in a project to modify Rb/Sr methods to date garnets and determine how long it took for them to grow (Christensen et al., 1988). This approach has many potential applications as are clearly enuciated by Rosenfeld et al. (1988) and may tell us when the first stage garnet actually grew.

Thompson et al. (1977) raised an important debate when they pointed out that unconformity textures could be produced in garnet porphyroblasts during a single prograde metamorphism. All that is required, according to their suggestion, is that an intermediate, garnet-consuming reaction interupt prograde garnet growth long enough during deformation for the matrix fabric to rotate relative to the porphyroblast. Renewed garnet growth, after the intermediate garnet-consuming reaction is no longer operating, would produce an outer shell of garnet with inclusion trails oriented at a high angle to inclusion trails in the inner garnet shell. As an example, Thompson et al. (1977) suggested that the breaking of the garnet-chlorite tie line to produce biotite and staurolite could interupt garnet growth. After consumption of chlorite by this reaction, garnet could grow again, perhaps by a continuous reaction which consumes biotite and staurolite and produces garnet. Downie (1982) extended this suggestion to include possible reactions involving non-AFM phases such as rutile.

This is an attractive alternative hypothesis to explain the unconformity textures. However, Cook and Karabinos (1988) emphasize that the unconformity texture is common throughout southeastern Vermont over a wide range of metamorphic grade and in a variety of bulk compositions at any given metamorphic grade. The timing of garnet growth with respect to deformation fabrics is also surprizingly consistent throughout southeastern Vermont. These observations are most easily explained by a change in the physical conditions of the first prograde metamorphism; it would require an amazing coincidence for prograde garnet-consuming reactions to commence in a wide variety of mineral assemblages at approximately the same time with respect to deformation.

Mafic schists: The petrology of mafic rocks at Stops 5 and 6 and along the route of field trip B-6, this volume, led by Rosenfeld et al. (1988) is presented by Laird and Albee (1981) and Boxwell and Laird (1987). Between Stops 1 and 3 (trip B-6), mafic schist changes from the epidote-amphibolite facies to the low-grade amphibolite facies, about 500 to 550°C based on garnet-biotite and calcite-dolomite geothermometry (Laird and Albee, 1981, Table 2). The change is mapped at the oligoclase isograd which is in exactly the same place for mafic rocks as pelitic rocks (mileage 4.0, Rosenfeld et al., this volume). Both albite and oligoclase occur locally above (to the south and higher grade than) the isograd. When/how did the oligoclase isograd get here from the Connecticut Valley trough (where it is mapped in mafic rocks by Mimi Boxwell and shown in Boxwell and Laird, 1987, Fig. 2)? Or did it? Near Jamaica the oligoclase isograd occurs in mafic rocks between stations 124 (with hornblende + albite) and 1001 (with hornblende + oligoclase). (See fig. 3 for sample localities.)

Along the traverse described by Rosenfeld et al. (1988, Stops 1 to 3), the change from titanite to rutile and/or ilmenite occurs within mafic rocks at about the same place as the oligoclase isograd. Hornblende is stable

above and below this isograd, indicating medium-pressure facies series metamorphism. Chlorite and epidote decrease in mode southward, while amphibole increases in mode, consistent with increasing metamorphic grade. A total fusion 40Ar/39Ar amphibole age at South Windham is 376 ± 5 Ma (Laird et al., 1984, sample V107A). At West Townshend a 40Ar/39Ar age spectrum on amphibole is saddle-shaped with the "bottom of the saddle" at about 377 ± 2 Ma (Laird and Sutter, unpublished data): (See Stop 5 for further isotopic data.)

Zoned amphibole with actinolite cores and hornblende rims occurs up to Stop 3 (our Stop 5). Locally, complexly zoned amphibole (anhedral hornblende overgrown by subhedral actinolite overgrown by hornblende, Plate 2a, Laird and Albee, 1981) is interpreted to have formed by polymetamorphism. Is the hornblende core Taconian and the retrotression implied by change of amphibole from hornblende to actinolite time-equivalent with the retrogression observed in the unconformity garnets? A zone of depletion between actinolite core and hornblende rim accompanied by smaller grains of amphibole is interpreted as a hiatus, between Rosenfeld's Acadian events I and II? Alternatively, could both hornblende core and actinolite core be pre-Acadian?

Zoned amphibole with actinolite cores and hornblende rims also occur in low-grade amphibolite from the Hoosac Formation (Turkey Mountain amphibolite) at stations 124 and 1001 (Stop 3) shown of Figure 3. Coexisting amphibole and plagioclase compositions indicate medium-pressure facies series metamorphism. between the epidote-amphibolite (124) and amphibolite (1001) facies. Cores are mottled as seen in backscattered electron image and locally show symplectic textures optically. Both samples are along thrust zones.

Zoned amphibole described above is within the garnet zone. Amphibole is not extensively zoned at high garnet grade (Stop 5) or staurolite-kyanite grade (Stop 6). Compared to the low and middle garnet zone, higher grade mafic schist contains somewhat more anorthitic plagioclase (oligoclase to andesine), amphibole with a bit more Al, Na, and K, biotite with more Al(VI), and chlorite with more Al(IV). Evidence of polymetamorphism within the high-grade mafic schists is seen at Stop 5 where amphibole is locally pseudomorphed by biotite, chlorite, plagioclase, epidote, quartz, and ilmenite/hematite. However, amphibole is not pseudomorphed in most layers, and plagioclase is zoned up grade (toward more anorthitic rims).

## FUTURE WORK

Dating of movement on thrust faults may be effected by obtaining absolute ages on minerals in fault zones and on "both sides" of the fault. Stratigraphy, structure, and petrology also hold keys for comparing relative metamorphic history across a fault zone.

Dating of garnets (Rosenfeld et al., 1988) are providing clues to time and duration of garnet growth. 40Ar/39Ar age spectra on unzoned hornblende give Acadian ages (if the spectrum is concordant) or saddle-shaped spectra (geologic meaning of extraneous Ar?). With a bit of luck laser studies of zoned amphibole will give metamorphic ages for the different compositional zones (Laird and Sutter, in progress).

Correlating deformation fabrics with map-scale structures and metamorphic minerals is extremely important. Rosenfeld (1968) and Rosenfeld et al. (1988) have showed how to use garnet for forensic studies. Will suggestions made herein for correlations between the geologic history of pelitic and mafic rocks "hold up" with further testing? Can the metamorphic and deformational histories of mafic rocks east and west of the Chester and Athens domes be "tied together"? Can one follow the oligoclase isograd in the Connecticut Valley trough into this isograd in pre-Silurian rocks? If so how is it related to the various deformation events suspected?

A big-picture question addressed by Stanley and Ratcliffe (1985) is why is there no evidence for high-pressure metamorphism in southeastern Vermont while the same units mapped in northern Vermont show medium-high and high-pressure facies series metamorphism (Laird and Albee, 1981)? Does this change really occur at about the present latitude of Rochester, Vermont, and if so, why?

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## ITINERARY

Assembly point is in front of U.S. Post Office in Jamaica, VT along Routes 30 and 100.

Mileage

0.0 Drive west on Route 30 and north on Route 100

1.4 Turn right onto Ball Mountain Dam Access Road

2.6 **STOP 1.** Park on left in grassy area at east end of road cut. Walk back to west to see rocks of the Middle Proterozoic Mount Holly Complex, which are here dominated by mafic to intermediate meta-igneous gneisses. The strong layering suggests that they may be meta-volcanic rocks. Contact between basement and cover rocks is east of long road cut and covered by grass. Outcrop of Tyson Formation containing quartz pebbles is first exposure of cover rocks east of basement. Here the Tyson Formation is a quartz, plagioclase, muscovite, chlorite, biotite schist with some pebbly layers.

Continue straight ahead.

3.1 End of road, go around circle.

3.3 **STOP 2.** Park on right side of road. At the west end of outcrop is a great exposure of plagioclase porphyroblast schist of the Hoosac Formation. At the east end of the outcrop is another great exposure of the high-alumina, chloritoid-paragonite schist of the Hoosac Formation, which is very similar to the Gassetts schist studied by Thompson et al. (1977). Mineral assemblages are also similar to those reported by Albee (1965) in central Vermont and by Cheney (1980) in western Massachusetts.

The contact between these two lithologies is very gradational. Both lithologies contain quartz, muscovite, chlorite, garnet, ilmenite, and epidote. The main difference between them is that the former contains the pair plagioclase-biotite and the latter contains paragonite-chloritoid. Rarely, three of these four minerals are found in the same rock. Note also the carbonate-rich pods and layers in the schist. Sphalerite is present but not common, and it may have been the source of Zn for staurolite when it appeared in higher-grade rocks. Garnets in the high-alumina schist display a textural unconformity, but it is rather subtle in many of the samples.

Return to Route 30.

4.85 Turn left, south onto Route 30 and head back to Jamaica.

6.3 Jamaica, U.S. Post Office.

8.3 Bridge over West River. High-alumina member of Hoosac Formation on left contains staurolite.

9.2 Turn left onto Turkey Mountain Road (gravel). Turkey Mountain Brook on right.

10.2 **STOP 3.** (Optional stop if weather, time, and enthusiasm permit.) Location PK1001, Fig. 3. Park along road. Cross bridge and head east upslope to top of ridge. Traverse upslope goes through the sequence: plagioclase porphyroblast schist, chloritoid schist (with staurolite), plagioclase porphyroblast schist, basement gneisses containing strong deformation fabrics, more plagioclase schist, and some mafic schist layers near the top of the ridge. Karabinos (1984a) explained this sequence by thrusting. The basement gneisses appear to be contained in fault bound slivers or horses (Fig. 3).

The amphibolite is coarse grained and amphibole is zoned with hornblende rims and actinolite cores. Plagioclase is oligoclase.

11.0 Bear right across bridge.

11.5 **STOP 4.** Park in grassy area on left. Walk north on road about 50 m and turn right, east, upslope to ledges. Very good outcrop of high-alumina schist member of the Hoosac Formation (station 120, fig. 3). Large garnets with well developed unconformity textures (this is the site of sample 120D, fig. 2, discussed by Karabinos, 1984a,b). Staurolite is present in some layers.

Turn around and return to Route 30.

13.7 Turn left onto Route 30.

14.0 Intersection with Route 100, continue straight on Route 30.

14.1 Outcrop of basement gneisses on left.

16.4 Outcrop on both side of road of the Moretown Member of the Missisquoi Formation.

17.9 **STOP 5.** Park on right side of road in lot overlooking Townsend Dam. A good place for lunch and a good place to ponder stratigraphy, structure, and petrology.

Known: 1) The west end of the roadcut is mapped as the Moretown Member of the Missisquoi Formation and is composed primarily of light gray, medium-grained quartz + oligoclase + biotite + white mica + chlorite + epidote + garnet + hematite/ilmenite schist with fascicles of hornblende pseudomorphed by biotite + chlorite + epidote + plagioclase + ilmenite/hematite and with accessory apatite, allanite, tourmaline, chalcopyrite, pyrite, and magnetite. Fine-grained white mica in the light gray schist looks like paragonite, but only muscovite analyses have been obtained by J.L. Interlayers of dark gray, medium-grained amphibolite are folded and boudinaged locally. They are composed of hornblende + epidote + biotite + chlorite + quartz + plagioclase + calcite + ilmenite with accessory white mica, apatite, pyrite, and chalcopyrite. Plagioclase is zoned outward from oligoclase to andesine.

2) Farther east are two layers (15 feet and 54 feet wide) of dark green to gray amphibolite (amphibole + epidote + quartz + plagioclase + carbonate + chlorite + ilmenite/hematite + garnet + biotite separated by light gray felsic schist (23 feet wide). These are the two amphibolite layers that are readily seen from across the street. Amphibolite interlayers occur within the felsic layer near the contacts with both amphibolites. Amygdaloidal-looking "enclaves" within the easternmost amphibolite are composed of plagioclase + quartz + minor amphibole.

3) The eastern part of the outcrop is composed primarily of a brown-weathered, light gray white mica + quartz + biotite + plagioclase + chlorite + garnet schist with rolled garnets. The rolled garnets give the rock a knotty appearance and prove to be excellent for forensic studies (Rosenfeld et al., 1988). Amphibolite occurs within the felsic layer, and the east end of the outcrop is amphibolite with cross-folial chlorite (amphibole + epidote + oligoclase + quartz + chlorite + biotite + ilmenite, rutile, pyrite, chalcopyrite, and apatite).

4) Amphibole from coarse-grained garnet amphibolite gives a concordant $^{40}Ar/^{39}Ar$ age spectrum with an age of 380 $\pm$ 2 Ma (Laird and Sutter, unpublished). Amphibole from medium-grained amphibolite gives a saddle-shaped spectrum with the "bottom of the saddle" at 389 $\pm$ 2 Ma. These data are consistent with isotopic studies presented by Rosenfeld et al. (1988) from this outcrop.

Questions: What is/are the nature of the contacts? If the west end of the outcrop is Missisquoi Formation and the thick felsic schist with rolled garnets on the east side of the outcrop belongs to the Ottauquechee Formation, there must be a fault and/or unconformity here somewhere. "Shredded" amphibolite in the two, thick amphibolite layers in the middle of the outcrop may mark faults. Yet presence of amphibolite layers within felsic layers near contacts between the two rock types suggest gradational contacts. Why does one see retrograded amphibole at the west end of the outcrop, but amphibole is "reasonably" fresh elsewhere? Cross folial chlorite and biotite altered to chlorite is seen throughout the outcrop.

Our route now follows that of Rosenfeld et al. (1988, mileage 13.1 to 21.6). Please consult that guide for more detailed discussion of the geology between Stops 5 and 6.

18.3 Cover rocks of the Hoosac Formation exposed on left side of road.

18.4 Covered bridge on right.

18.7 Sheared augen gneiss of the basement on left.

20.0 Turn left, north, onto Route 35 in Townsend, VT.

23.4 Bear left on road to Grafton.

25.2 Turn sharply left onto road heading uphill.

25.4 **STOP 6**. Park on left side of road. Very sheared augen gneiss mapped by Doll et al. (1961) as part of the Cambrian (?) Cavendish Formation. Karabinos and Aleinikoff (1988) dated zircons from this rock using the U-Pb method that gave an upper intercept age of 955 + 5 Ma. They interpreted this as the age of crystallization which indicates that this augen gneiss is not part of the Late Proterozoic to Cambrian cover sequence. It seems that we must resort to some structural explanation for the occurrence of cover rocks between the Bull Hill augen gneiss and other gneisses of the Mount Holly Complex. An interesting feature of the geochronology is that the lower intercept age suggests modern lead loss and does not record any trace of either the Taconic or Acadian orogenies.

Continue straight ahead on dirt road. In only a mile we cross the Pinney Hollow Formation at Rosenfeld's unconformity garnet (Rosenfeld et al., 1988, Stop 6) and into the Moretown Member of the Missisquoi Formation. This mile then covers the stratigraphy seen from Stops 1 to 5.

26.4 **STOP 7**. Park on dirt road along Power line. Outcrops along the power line of felsic rock with mafic interlayers very similar to the west end of Stop 5. The light gray layers are composed of: quartz + plagioclase + white mica + biotite + epidote (with allanite) + garnet + opaques + apatite). The dark-gray layers are composed of hornblende + oligoclase + quartz + garnet + epidote + biotite + chlorite (late?) +

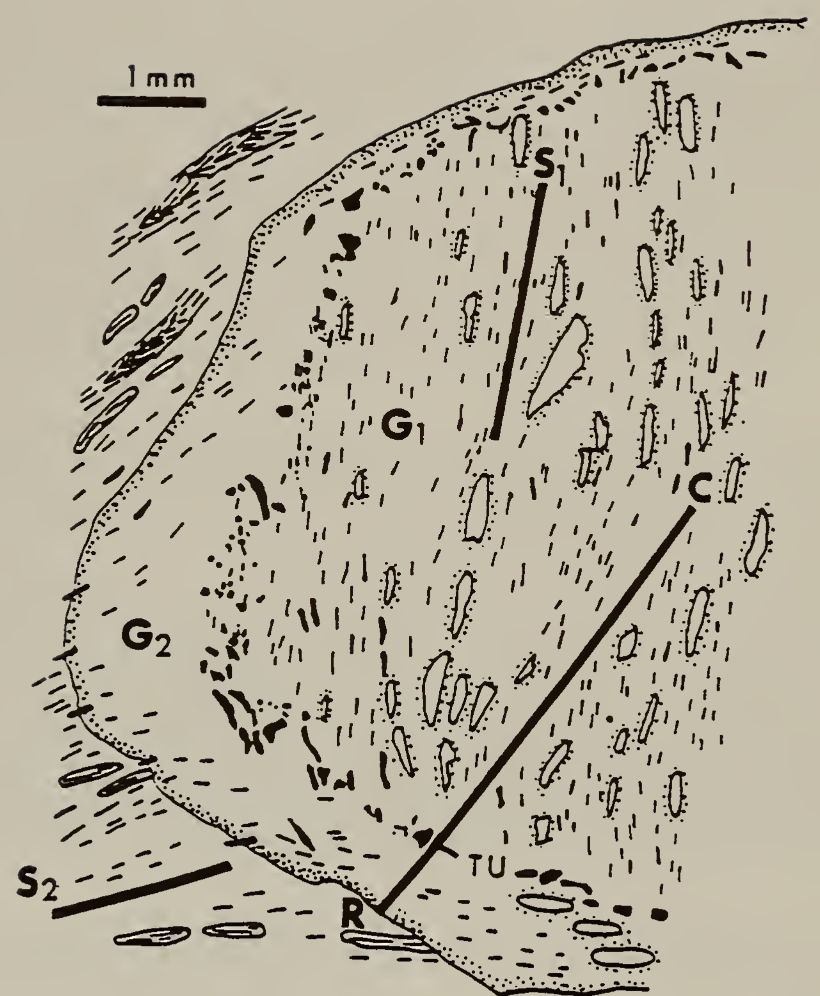
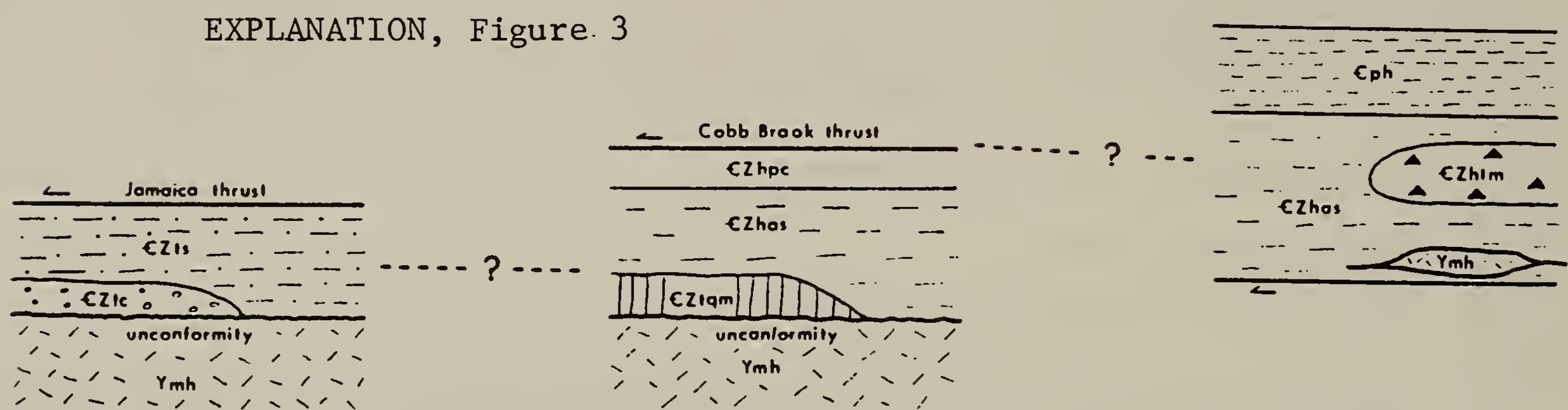



Figure 2: Drawing of garnet porphyroblast from Stop 4 (loc. 120, Fig. 3) showing early garnet growth (G1) separated from later garnet growth (G2) by a textural unconformity (TU). S1 defined by inclusions within garnet core is truncated by S2 within matrix and garnet rim along TU. Inclusions are chloritoid, white mica, and ilmenite. Analytical data along core (C) to rim (R) traverse are presented by Karabinos (1984a, Fig. 10).

EXPLANATION, Figure 3




Western Sequence

| | | |
|---|---|---|
| Tyson Formation | Qurtz-muscovite schist | €Zts |
| | Conglomerate | €Ztc |
| Mount Holly Complex | | Ymh |

Central Sequence

| | | |
|---|---|---|
| Hoosac Formation | Paragonite-chloritoid schist | €Zhpc |
| | Albite porphyroblast schist | €Zhas |
| Tyson Formation | Quartzite-dolomite unit | €Ztqm |
| Mount Holly Complex | | Ymh |

Eastern Sequence

| | | |
|---|---|---|
| Pinney Hollow Formation | | €ph |
| Hoosac Formation | Turkey Mountain Member | €Zhtm |
| | Albite porphyroblast schist | €Zhas |
| Mount Holly Complex | | Ymh |

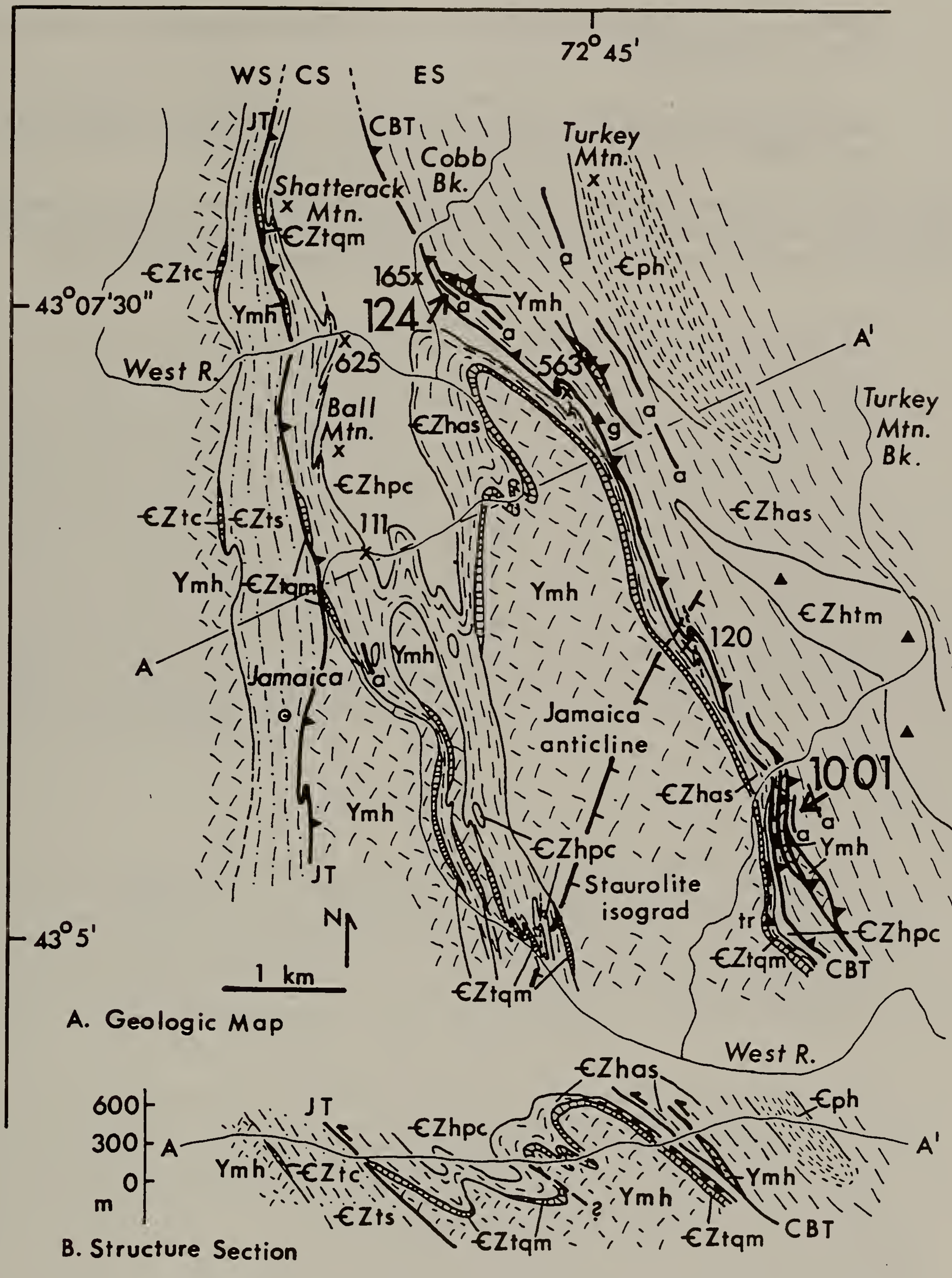



Figure 3: Geologic map and cross section (no vertical exaggeration) of the Jamaica area from Karabinos (1984a). Stops 1 and 2 (traverse west of Ball Mtn.); Stop 3 (traverse to locality 1001), Stop 4 (locality 120). Western sequence (WS) is separated from the Central Sequence (CS) by the Jamaica thrust (JT), and CS is separated from the Eastern Sequence (ES) by the Cobb Brook thrust (CBT). Explanation on previous page.

ilmenite/hematite + apatite + pyrite + chalcopyrite. The metamorphic grade is just a tad higher (staurolite-kyanite zone) that at Stop 5.

Back down the road toward Rosenfeld et al.'s (1988) Stop 6 and on the south side of the road in the woods is an outcrop of amphibole + biotite + plagioclase + quartz + chlorite + epidote + garnet + opaque + apatite schist. Is this the same as one of the thicker amphibolites at Stop 5?

Return to Hwy 35. Brattleboro, Interstate 91, and highway 9 are reached by turning right on highway 35 and then left on highway 30 at Townshend. If you are northward bound, turn left on highway 35 to Chester, where highway 11 goes east to Interstate 91 and route 103 takes you north to the Gassetts schist and highway 100.

# ROOT ZONE OF THE BERNARDSTON NAPPE AND THE BRENNAN HILL THRUST INVOLUTED BY BACKFOLDS AND GNEISS DOMES IN THE MOUNT GRACE AREA, NORTH-CENTRAL MASSACHUSETTS

by
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## PURPOSE OF TRIP

A new stratigraphic and structural interpretation of the Mount Monadnock area, New Hampshire, (P. J. Thompson, 1985 and this volume) and new work by Elbert (1984, 1986, 1987, and this volume) in the Bernardston area, Massachusetts and Hinsdale area, New Hampshire, has pointed the way to yet another reinterpretation of the complex geology of the Mount Grace area. After early work by B. K. Emerson (1898, 1917), Jarvis B. Hadley (1949) mapped the area and was the first to introduce M. P. Billings' (1937) western New Hampshire stratigraphy to Massachusetts. Robinson (1963) found far greater structural complexity than envisioned by Hadley and showed that much of what Hadley had mapped as Lower Devonian Littleton Formation should be assigned to the Middle Ordovician Partridge Formation. In considering a synthesis of the regional fold nappes (Thompson et al., 1968) extensive remapping was done (Robinson, 1967; 1977) bringing the map pattern much closer to the one currently shown (see also Huntington, 1975). The present round of revisions, spurred by the work of P. J. Thompson and D. C. Elbert, was occasioned by the following new factors: 1) The virtual certainty that the gray feldspathic schists previously assigned to an eastern facies of the Devonian Littleton Formation are actually Lower Silurian Rangeley Formation. 2) The iron formation previously assigned to the eastern Littleton should be assigned to the middle Silurian Perry Mountain Formation. 3) Both of these rock types belong to the Monadnock sequence of strata and were transported into juxtaposition with the traditional Bronson Hill sequence by the nappe-stage Brennan Hill thrust.

## STRATIGRAPHY

The strata of the region may be divided into four major groups as follows: 1) Gneisses and related rocks exposed in cores of domes that are in the age range Late Precambrian-Ordovician; 2) Middle Ordovician Ammonoosuc Volcanics and Partridge Formation; 3) Silurian-Lower Devonian stratified rocks of the Connecticut Valley and Merrimack belts; and 4)Triassic-Jurassic sedimentary rocks and basalts of the Connecticut Valley Mesozoic basins. These are cut by a variety of pre-tectonic, syn-tectonic, and post-tectonic Silurian-Devonian intrusions ranging from gabbro to granite, and by Jurassic and Cretaceous diabase dikes. General stratigraphic relations and stratigraphic problems have been covered extensively elsewhere and only features related to the present work are summarized here.

The origin and age of pre-Middle Ordovician basement rocks in the gneiss domes continues to be a problem (Robinson, 1981). The stratified gneisses in the core of the Pelham dome (Figure 1) have been considered Late Precambrian on the basis of one zircon age (Naylor et al., 1973, Zartman and Naylor, 1984) and lithic and geochemical similarities to rocks in southeastern Connecticut. This age is confirmed in a new preliminary zircon age of 606 m.y. on the Dry Hill Gneiss by R. D. Tucker (pers. comm. 1988). Hodgkins (1983, 1985) has shown that the dominant felsic gneisses of this sequence,with low normative anorthite and extremely low MgO/(MgO+FeO) ratios, have a major and trace element chemistry consistent with interpretation as a sequence of chemically evolved alkali rhyolites that might have erupted in a rifting environment. The relics of an apparently pre-Devonian granulite-facies metamorphism that survived Acadian kyanite-muscovite overprinting (Robinson, Tracy and Ashwal, 1975; Roll, 1986, 1987) are still of unknown age, but were tentatively assigned to the Late Precambrian (Robinson, 1983). However, preliminary late Devonian U-Pb monazite and zircon ages from a sillimanite-orthoclase pegmatite within this schist raise serious doubts about this conclusion (R. D. Tucker, pers. comm., 1988). The layered plagioclase gneisses and amphibolites that are called Fourmile Gneiss where they overlie the Late Precambrian strata in the Pelham dome, and called Monson Gneiss in most other areas, have not yet been satisfactorily dated. The Monson Gneiss in Massachusetts has yielded a zircon age of about 450 m.y., but

the New London Gneiss of southeastern Connecticut, lithically identical to part of the Monson, gives an age over 500 m.y. (Zartman and Naylor, 1984). Earlier views that the Monson Gneiss might represent metamorphosed volcanics just slightly older than the overlying Ammonoosuc Volcanics now seem improbable on the basis of new field, petrologic, and geochemical studies (Robinson et al., 1986; Hollocher, 1987; Lent, 1987). These suggest that Monson Gneiss is dominantly a highly deformed plutonic complex, with a tonalitic matrix filled with inclusions of more mafic tonalite, gabbro, and even gabbroic anorthosite, and cut by mafic dikes metamorphosed to amphibolite. Supracrustal rocks, which may be xenoliths, are limited to a single outcrop area of calcareous quartzite and layered amphibolite blocks that may represent metamorphosed basaltic tuffs. Massive batholithic-looking gneisses in the domes, including the Pauchaug Gneiss of the Warwick dome, appear to give consistent isotopic ages around 450 m.y. (Zartman and Leo, 1984).

The Ammonoosuc Volcanics of presumed Middle Ordovician age, is the basal unit of the cover sequence and its detailed stratigraphy is crucial to understanding the basement-cover relationship. In the early 1970's a basal quartzite and conglomerate lens was found by Robinson where the Ammonoosuc overlies the Monson Gneiss in the Orange quadrangle, and in 1983 thin lenses of quartzite were found precisely on the same contact in two localities in the Quabbin Reservoir area. Despite numerous suggestions that the massive batholithic-looking gneisses are intrusive into the Ammonoosuc (Leo, 1985) and the radiometric dating that appears to make this permissible (Zartman and Leo, 1984), there remains no good documentation of intrusive gneisses substantially truncating a well defined Ammonoosuc stratigraphic sequence, and the balance of field evidence indicates the gneisses are older than Middle Ordovician and possibly as old as late Precambrian. R.D. Tucker of Toronto University (pers. comm. 1988) has recently produced new zircon data based on several carefully selected fractions of a sample of Monson Gneiss from the Quabbin Reservoir area. This indicates an age on Concordia of 435 m.y., which is early Silurian, and adds further vexation to a vexatious problem.

The research of Schumacher (1983, Schumacher and Robinson, 1986, 1987; Schumacher, 1988) and Hollocher (1983, 1985, Robinson et al., 1986) shows that the volcanics of the Ammonoosuc and the overlying Middle Ordovician Partridge Formation are quite similar in their major and trace elements to the low-K tholeiites, andesites, and dacites, as well as K-bearing rhyolites of modern island arcs such as Tonga and New Britain. An important aspect of their conclusions is that the K-bearing peraluminous rhyolites that characterize the upper part of the Ammonoosuc and continue into the Partridge, are not the product of melting of subducted North American continental crust as suggested by Robinson and Hall (1980), nor melting of subducted pelitic sediments, but were produced by melting of amphibolite or granulite of tholeiitic basalt composition within the magmatic arc complex. A curious problem is that the presumed medial Ordovician age of these supposed arc volcanics is the same as the time of emplacement of the Giddings Brook slice of the Taconic allochthon. This was not when Iapetus was beginning to close but essentially when it had virtually completed closing. A possible explanation for the time lag is offered by new models for subduction zone magmatism (Kushiro, 1987) indicating that most arc basaltic magmas are derived by melting of upper plate mantle when influenced by aqueous fluid diffusing slowly upward from the subduction zone.

Stratified Silurian-Devonian rocks occur in two main belts, the Connecticut Valley belt to the west and the Merrimack belt to the east (Zen et al., 1983). In the Connecticut Valley belt there are a host of stratigraphic problems, including the recent finds of both Middle to Late Ordovician graptolites (Bothner and Finney, 1986) and late Lower Devonian (Emsian) plant fossils in similar rocks (N.L. Hatch, pers. comm., 1986). These problems are by no means even partially resolved by the proposed Whately thrust (Robinson, Hatch and Stanley, 1984, and in press) that tentatively solves the stratigraphic dilemma of the Devonian(?) Erving Formation in the western part of the Orange area.

Along the eastern edge of the Connecticut Valley belt, there has been a significant advance in the paleontological control of the thin strata characteristic of the Bronson Hill anticlinorium. Specifically, Elbert, et al., 1988; and Elbert this volume, have identified a rich conodont fauna in garnet-grade marble of the Fitch Formation in the inverted limb of the Bernardston nappe at Bernardston. This marble lens lies stratigraphically above, though structurally below, fossiliferous Clough Quartzite containing a poor fauna suggesting Lower Silurian age (Boucot et al., 1958). The marble lies structurally above, though stratigraphically below, gray garnet phyllites of the Devonian Littleton Formation, and appears to be truncated laterally by an unconformity at the base of the Littleton which is more generally in direct contact

with the Clough. The conodont fauna gives a strong indication that the Fitch marbles in this location are earliest Devonian and not middle or upper Silurian as has been found elsewhere.

A key stratigraphic feature for the present studies is the eastward thickening of the Silurian sequence from the thin section of Clough Quartzite and local Fitch Formation in the Bronson Hill anticlinorium, at the east edge of the Connecticut Valley belt, into the thick sequence of the Merrimack belt. This consists of Rangeley Formation (lower Silurian), Perry Mountain Formation (middle Silurian), Francestown Formation (middle Silurian), and Warner Formation (upper Silurian) as defined by Peter Thompson (1985) in the Monadnock area (Figure 1), and correlated in detail by him with Silurian sequences in central New Hampshire and northwestern Maine (Hatch, Moench, and Lyons, 1983). It is the distinctive character of the well defined Monadnock sequence and its differences with the thin Bronson Hill sequence that makes possible the mapping of the early west-directed thrust nappes that have become such an important part of recent tectonic reconstructions. One of these thrust-nappes carried rocks of the Monadnock sequence westward into the Hinsdale, N.H. area (Figure 1), where Elbert has discovered a distinctive horizon of garnet quartzite, and magnetite-cummingtonite iron formation at the top of the middle Silurian Perry Mountain Formation. This has made possible the new interpretation of the Mt. Grace area, Massachusetts (Figure 1) covered in this field trip, where identical rocks had previously been assigned to the Lower Devonian Littleton Formation (Huntington, 1975).The Monadnock sequence appears to extend from southwestern New Hampshire across Massachusetts in a very tight zone east of the main body of Monson Gneiss and west of the Coys Hill pluton (Figure 1). In Massachusetts it has the distinctive graphite-pyrrhotite calc-silicate rocks of the middle Silurian Francestown Formation, but generally lacks the upper Silurian Warner Formation. Stratigraphy still provides the essential control for local and regional structural and tectonic studies.

## STRUCTURAL GEOLOGY

Acadian deformations in central Massachusetts and adjacent New Hampshire have been summarized elsewhere (Robinson, 1979; Robinson and Hall, 1979; Hall and Robinson, 1982). These are broadly divided into an early nappe stage, an intermediate backfold stage, and a late gneiss dome stage, each with many complications.

Recent research in the Monadnock area, New Hampshire (P.J. Thompson, 1985), the Hinsdale area, New Hampshire (Elbert, 1986), the Mt. Grace area, Massachusetts (Robinson, 1987), and the Sturbridge area, Massachusetts and Connecticut (Berry, 1987a, 1987b) has shown that the early west-directed fold nappes are truncated by a slightly later set of west-directed thrust nappes. In southwestern New Hampshire and adjacent Massachusetts the two major nappe-stage thrust faults are: 1)The Brennan Hill thrust carrying previously folded Monadnock sequence strata across Bronson Hill sequence strata containing the Bernardston fold nappe. 2) The Chesham Pond thrust carrying Rangeley Formation and Kinsman Granite over previously folded rocks of the Monadnock sequence (Figures 2, 3, ). The Brennan Hill thrust appears to trace southward from the Monadnock area through the Mt. Grace area and along the east margin of the main body of Monson Gneiss to Connecticut, where Pease (1982) had previously identified the Bone Mill Brook fault zone . The Chesham Pond thrust appears to trace southward from the Monadnock area along the west margin of the Coys Hill pluton and also may extend into Connecticut.

The structural features tentatively assigned to the backfold stage include the following, not necessarily in chronological order: Evidence for the first three of these is found in the Mount Grace area.

1) Longitudinal flowage of the main and Tully bodies of Monson Gneiss, from a position south of Quabbin Reservoir, 50 km northward to a position overlying strata in the Mt. Grace area. This longitudinal flowage caused recumbent folding that involuted the axial surfaces of early fold nappes as well as the trace of the Brennan Hill thrust.

2) East-southeast directed recumbent folding of the basement-cover contact in the Pelham gneiss dome (Ashenden, 1973) and the Keene gneiss dome (Robinson, 1963, 1967; Schumacher and Robinson, 1986). Amphibole lineation associated with this phase has been identified locally in the Keene gneiss dome and in gedrite gneisses overgrown by cordierite believed to have been produced by

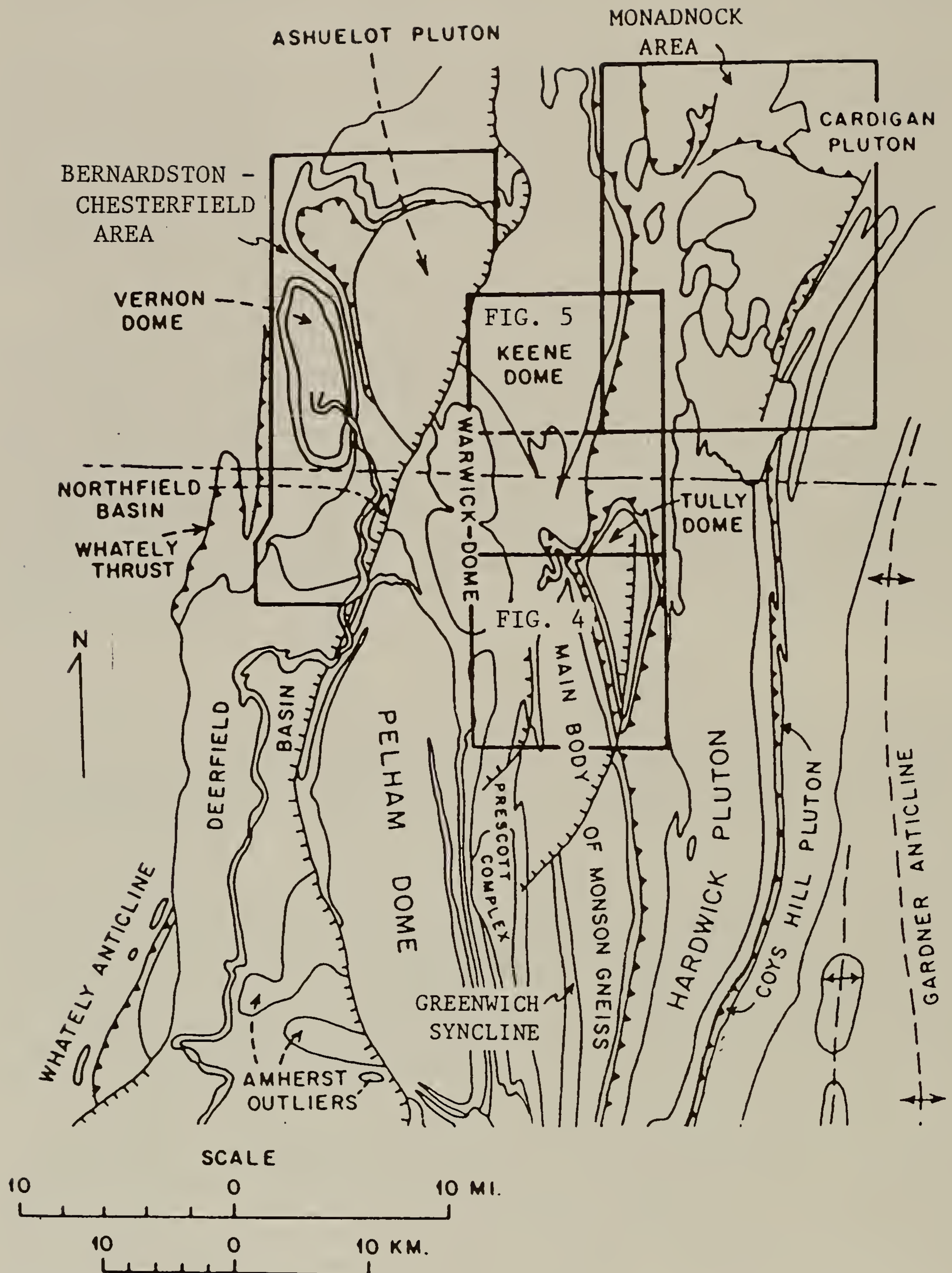



Figure 1. Geologic index map of north-central Massachusetts and adjacent New Hampshire and Vermont showing Mount Grace area (Figures 4 , 5) in relation to Monadnock and Bernardston-Chesterfield areas.

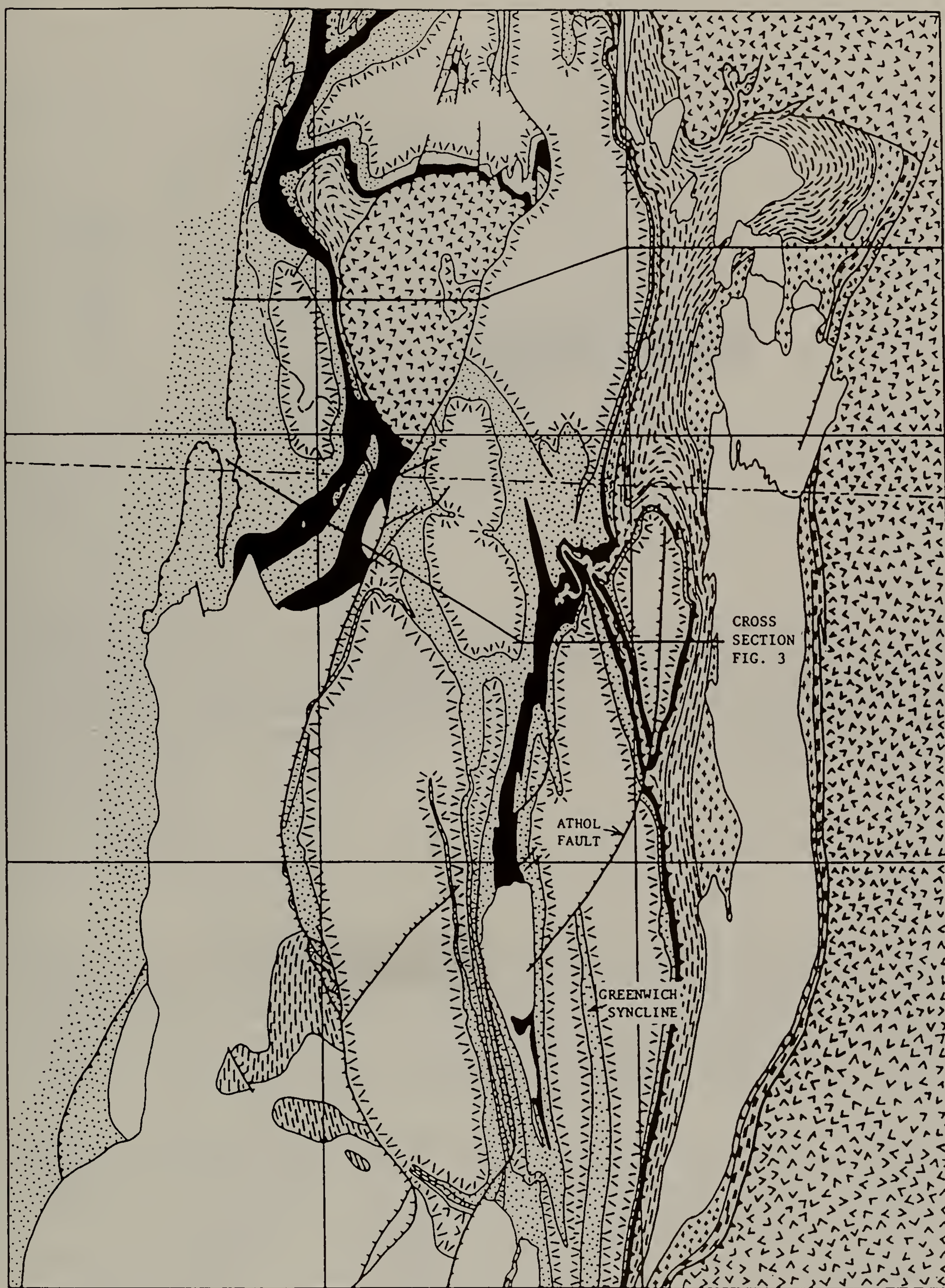



Figure 2. Map of north-central Massachusetts, southwestern New Hampshire, and adjacent Vermont showing distribution of three tectonic levels separated by thrust faults (with teeth). Within the three thrust sheets patterns indicate structural position on earlier recumbent folds. Mesozoic faults are hatchured on downthrown blocks; intrusive rocks and Mesozoic strata are unpatterened. Location of cross section in Figure 3 is also shown in northern Massachusetts. Approximately E-W line in southern New Hampshire is location of cross section of Elbert (this volume). Key to symbols is shown in Fig. 2A.

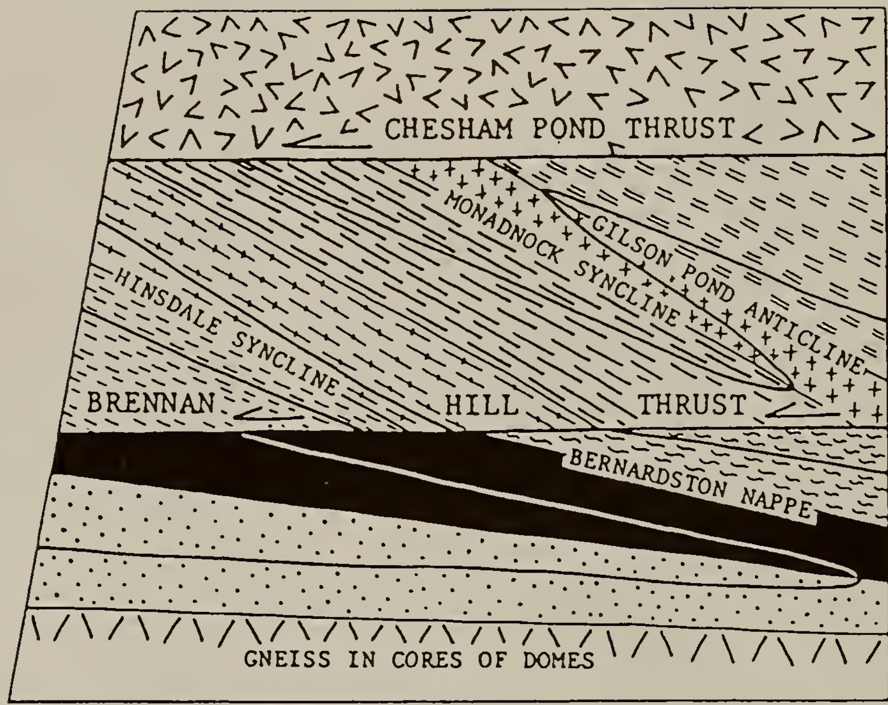



Figure 2A. Schematic cross section showing three tectonic levels separated by thrust faults. At the lowest level, stratigraphy of the Bronson Hill sequence, folded by the Bernardston nappe, is truncated upward by the Brennan hill thrust. At the middle level, the Monadnock sequence, folded by the Monadnock syncline and related recumbent folds, is truncated below by the Brennan Hill thrust and above by the Chesham Pond thrust. At the highest level, the Kinsman Granite and associated Rangeley Formation are truncated below by the Chesham Pond thrust.

tectonic unloading during the gneiss-dome stage. At the south end of the Pelham gneiss dome these structural features are truncated by the Belchertown Quartz Monzodiorite intrusion that has a zircon age of 380 million years (Ashwal et al., 1979). The rocks of both the dome and the intrusion are structurally overprinted by the dome stage of deformation. Unfortunately no field relations have yet been found that show the relative age of these east-directed recumbent folds to the west-directed fold or thrust nappes, or to other features tentatively included in the phase of backfolding.

3) Eastward overturning of axial surfaces of west-directed fold nappes and west-directed thrust surfaces. Earlier it was hoped that major axial surfaces related to this backfolding could be identified in central Massachusetts but nothing conclusive has yet been found. P.J. Thompson (1985) and H.N. Berry (1987) have independently suggested the possibility of grand overturning of the entire eastern part of the orogen from the Monson Gneiss across the entire Merrimack belt and possibly beyond. Tentatively associated with this eastward overturning, but not conclusively linked with it, is an E-W trending linear fabric and E-W trending minor folds (Robinson, 1979, Peterson, 1984) that are progressively overprinted westward by north or northeast-trending folds and fabrics definitely related to the dome stage. The E-W trending fabrics appear both in metamorphosed sedimentary rocks and a wide variety of tonalitic through granitic intrusions.

4) Development of a series of mylonites in metamorphosed sedimentary and intrusive rocks. These mylonites contain an E-W trending lineation believed to be related to the shear direction and this lineation is parallel to the E-W lineation described under 3) above. The mylonites cut across coarse-grained migmatitic schists and gneisses formed during peak granulite-facies metamorphism. The

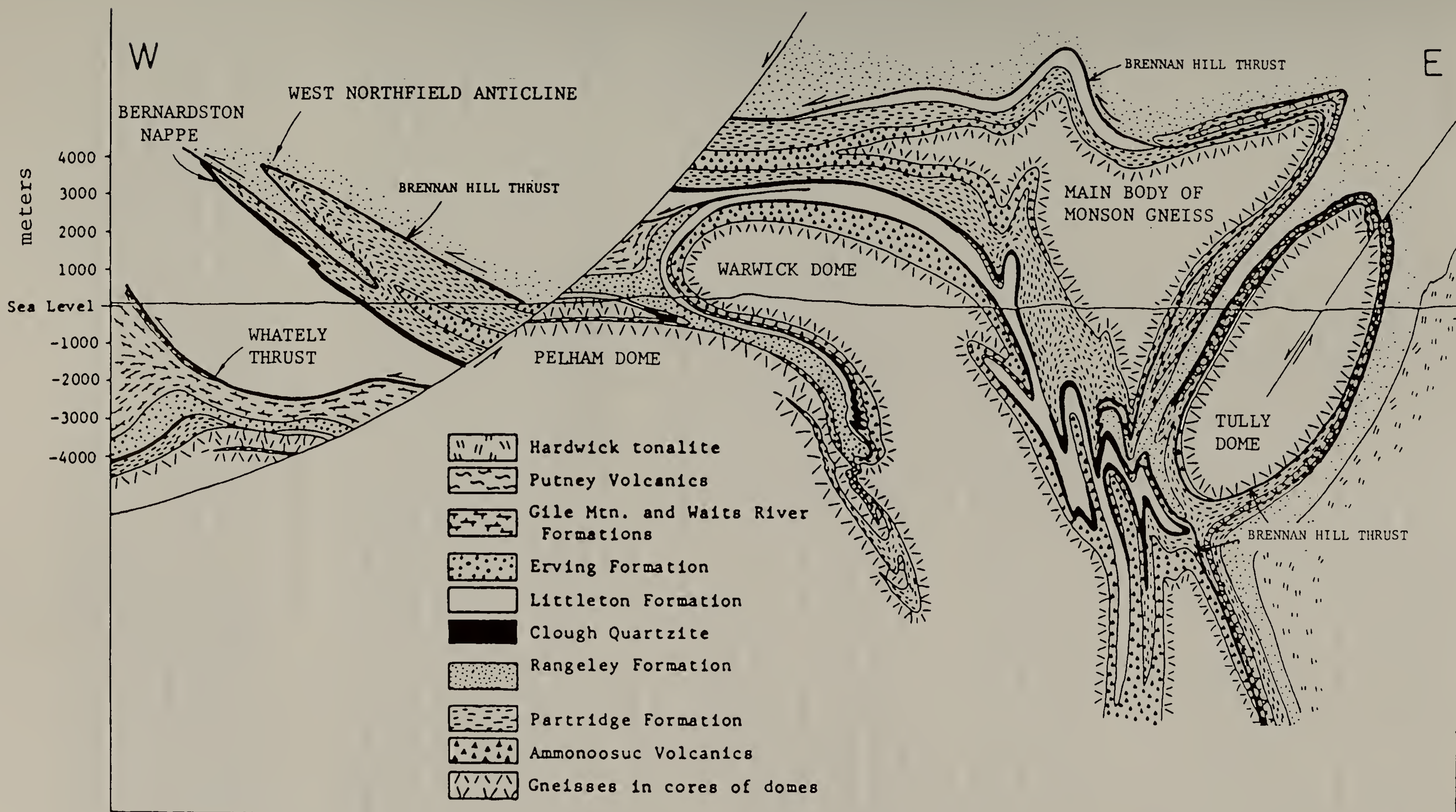



Figure 3. Cross section showing the probable stratigraphic and structural relationships between the Orange area (Robinson, 1977, 1987) and the Bernardston-Northfield area (Elbert, 1984). The Connecticut Valley border fault is shown in simplified form as a listric normal fault with a circular surface. The vertical component of the net slip is about 15,000 feet and the angle of rotation of the hanging wall block was about 16 degrees. For line of cross section see Figure 2.

mylonites are deformed by northeast-trending minor folds related to the dome stage. In an early structural correlation (Robinson, 1979, Robinson et al., 1982) it was proposed that the mylonites formed during late stages of backfolding following peak metamorphism. With discovery of early west-directed thrust nappes by Berry (Robinson et al., 1986) it was alternatively suggested that the mylonites are related to the thrust nappes which were then tentatively considered to post-date the peak granulite facies metamorphism. This alternative has now been definitively disproved for three reasons: a) Several thrust surfaces related to thrust nappes are truncated by tonalite intrusions (Berry, 1987a); b) Many tonalite intrusions are involved in the mylonites, some of them extensively so (Finkelstein, 1987; Berry, 1987b); and c) The shear sense of the mylonites studied so far, including mylonitized tonalites, indicate a consistent west-over-east shear sense inconsistent with the west-directed thrusting. It is thus now concluded that mylonite formation was part of the phase of backfolding as originally suggested, and unrelated to thrust nappes.

The swirling pattern of lineations related to the gneiss-dome stage of deformation, superimposed on all previous features, still needs to be studied further and traced from north-central Massachusetts to the Connecticut line. In the Pelham dome there is conclusive evidence that this lineation, trending N-S parallel to the dome axis, is parallel to the transport direction of a series of sheath folds (Ashenden, 1973, Onasch, 1973, Robinson, 1979). However, nearby detailed features in cover rocks (Michener, 1983) seem to deny this conclusion. The suspicion remains that the pattern of lineations associated with the dome stage, with its evidence of the gravitational rise of low density buoyant basement strata, may also reflect a deep-seated ductile regime of longitudinal shear related to the terminal phase of Acadian collisional tectonics (cf. Ellis and Watkinson, 1987).

The region was subjected to large-scale extensional faulting during Triassic-Jurassic time. Most important was the Connecticut Valley border fault that is thought to be listric in character. At the New Hampshire-Vermont line, reconstructed cross sections of pre-Mesozoic rocks suggest the west side was downthrown about 5 km (Figure 3) and indications are that displacement on the same fault at the Connecticut line may have been as much as 8 km. The chief importance to the present study is that the fault allows observation of the metamorphic rocks at widely different tectonic levels. In addition, more local faults with displacements of about 0.5 km are important in solutions of local structural problems. In particular, a southward extension of the Athol fault into the Quabbin Reservoir area is believed to cut across the Acadian Greenwich syncline (Figure 2). The location of the proposed fault is being traced by magnetic surveys on the ice of Quabbin Reservoir in winter. If the present prognosis is correct, the stratigraphic syncline is actually a structural anticline exposing younger rocks and is a faulted southward extension of the Walnut Hill anticline in the Orange area to the north.

A second indication of Mesozoic tectonic activity is a group of diabase dikes and sills, until recently all presumed to be early Jurassic. A study by McEnroe in the last year combining paleomagnetism, geochemistry, and K-Ar dating has come up with some surprising results (McEnroe and Brown, 1987; McEnroe et al., 1987; McEnroe, 1988). Several of the dikes and sills show 40 degree north normal inclinations consistent with a late Jurassic age. Two distinctive tholeiitic diabase dikes at Bliss Hill in the Mount Grace area (Stop 3, this field trip) give 60 degree north normal inclinations consistent with normal periods in the Cretaceous and have yielded a K-Ar whole rock age of 121 m.y. A system of tholeiitic olivine diabase sills in the Quabbin Reservoir area give 60 degree north reversed inclinations consistent with reversed periods in the Cretaceous and have yielded K-Ar ages as young as 119 m.y. The major- and trace-element geochemistry of these Cretaceous intrusions is completely different from the more abundant Jurassic intrusions in the region.

## METAMORPHISM

Details of the metamorphic zones in central Massachusetts, based on the petrology of pelitic schists, are given by Tracy et al., (1976), Tracy (1978), Robinson et al., (1982), and Robinson et al., (1986), although much detailed analytical data has yet to be published. Hollocher (1985) and Renate Schumacher (1986) have given details of progressive reactions in amphibolites in the same region, and Schumacher and Robinson (1986, 1987) have given details concerning the formation of cordierite during progressive unloading in the Keene gneiss dome. The relations of metamorphic zones and metamorphic reactions to the structural evolution of the region is still poorly understood despite an impressive amount of detail in specific locations. This is illustrated by details in three specific areas.

In the Hinsdale area, New Hampshire, it has long been known that the sillimanite isograd is concentric to the Ashuelot pluton of Kinsman Granite (Moore, 1950; Trask, 1964; Thompson et al., 1968). In an earlier tectonic model (Thompson et al., 1968) this was thought to be due to nappe-stage recumbent folding of the Kinsman, possibly still in a partially liquid state, to a position tectonically above the Bernardston nappe and the autochthonous Vernon dome. In the presently favored model (P.J. Thompson, 1985; Elbert, 1986; Robinson, 1987) the Kinsman of the Ashuelot pluton is considered to have been emplaced largely in a solid state by the Chesham Pond thrust, tectonically above the Brennan Hill thrust sheet which was itself tectonically emplaced across the previously formed Bernardston fold nappe and autochthonous rocks. Elbert (1987) has demonstrated significant differences in early growth history of garnets on opposite sides of the Brennan Hill thrust near Hinsdale, with later convergence to similar zoning paths in the outer rims. Despite this evidence that metamorphism was well underway prior to thrusting and that the sillimanite isograd is spatially related to the Chesham Pond thrust sheet, the isograd itself actually cuts far below the lower Brennan Hill thrust, through the axial surface of the Bernardston nappe and into autochthonous rocks of the east limb of the Vernon dome. Thus, the ultimate positioning of the sillimanite isograd must have taken place well after thrusting.

In the Ammonoosuc Volcanics at the south end of the Keene gneiss dome amphibolites and gedrite gneisses contain an amphibole lineation parallel to the axes of local backfold-stage southeast-directed recumbent folds. The gedrite lineation is overgrown by cordierite that forms complex reaction rims around sillimanite and other aluminous minerals as a result of unloading believed to have taken place during the rise of the gneiss dome. Recently Seifert and Schumacher (1986) have developed a pressure calibration based on the equilibrium between cordierite, zincian spinel, and quartz, that has been applied to the assemblages in these aluminous enclaves, and indicates pressures of 6.3 to 3.7 kbar for the time of enclave formation. Gedrite-garnet gneisses commonly contain both kyanite followed by sillimanite, and also conclusive evidence for kyanite forming as a retrograde product with chlorite and quartz after cordierite (Robinson, 1963; J.C. Schumacher, pers. comm. 1987). Garnet with up to 32% pyrope content in some of these rocks may be a relict of an earlier, higher pressure history before the cordierite-forming reactions (Schumacher and Robinson, 1986, Table G-8). Collecting in summer 1987 has led to discovery of previously unrecognized assemblages of coexisting Mn-rich orthopyroxene and augite in the middle Garnet-Amphibole Quartzite Member of the Ammonoosuc (J.C. Schumacher, pers. comm. 1987) where more normal Fe-Mg rich assemblages contain only amphiboles. This appears to be an example where abundant MnO stabilizes an assemblage normally considered characteristic of the granulite facies into P-T conditions of the amphibolite facies for more normal rock compositions.

The granulite facies area near Sturbridge, Massachusetts appears to be the area of most intense Acadian metamorphism in North America and its complex tectonic, mineralogic, and thermal evolution requires much more intense study. The widespread occurrence of sillimanite pseudomorphs after andalusite indicates an early low-pressure metamorphic history. That this low-pressure history involved high temperatures and partial melting is suggested by the occurrence of andalusite pseudomorphs in pegmatites and the widespread occurrence of cordierite-bearing pegmatites. In the one sample of cordierite pegmatite that has been studied in detail (Tracy and Dietsch, 1982) the cordierite has a 10% higher Fe/(Fe+Mg) ratio than typical cordierite in sillimanite-garnet-cordierite-quartz assemblages in the adjacent granulite facies pelites. This is taken to indicate that the partial melting that produced the pegmatite took place at considerably lower pressures than those indicated by peak granulite facies assemblages. Further, this particular cordierite is shown to have been in the process of breaking down on its own composition to an intergrowth of sillimanite, garnet, and quartz plus more Mg-rich cordierite. A pristine suite of 48 samples of such cordierite pegmatites has been collected for study to try to better understand the early history of partial melting in this region.

Peak-metamorphic sillimanite-garnet-cordierite-quartz-biotite assemblages in pelites in the region suggest pressures of 685-740 °C and pressures of 6.1 - 6.3 kbar. Hollocher (1985; Robinson et al., 1986) has reported several occurrences of the classic granulite facies assemblage of orthopyroxene-orthoclase-quartz (Figure 5) and several other occurrences where orthoclase adjacent to orthopyroxene has been replaced by a high temperature retrograde symplectite of Ti-rich biotite and quartz. Several well studied occurrences of the assemblage orthopyroxene-garnet-plagioclase-quartz (Hollocher, 1985; Robinson et al., 1986) can be applied to the pressure calibration of Perkins and Chipera (1985) and yield pressure estimates of 5.2 to 7.0 kbar at 700 °C.

The late metamorphic history of this region is also intriguing. Keys to this late history are found in assemblages developed in mylonites that cross cut the peak metamorphic fabrics. Many of the mylonitic rocks are not distinct petrologically from sheared versions of the peak metamorphic assemblages. However, a few have undergone such severe grain size reduction that they have been able to undergo complete recrystallization to new fine-grained assemblages representative of a different metamorphic facies (Robinson et al, 1977; 1982; 1986). A mylonite in one host rock consisting of coarse quartz, K-feldspar, sillimanite, cordierite and biotite has recrystallized to a new assemblage of quartz, K-feldspar, sillimanite(?), garnet, and Mg-richer biotite. Garnet-biotite Fe-Mg exchange thermometry suggests the mylonite recrystallization took place at around 550 °C and the Mg-rich garnet composition (approx. 30% pyrope) suggests crystallization at a minimum pressure of 7-8 kbar. If these indications are correct, then it would appear that the rocks of the central Merrimack belt may have continued to be compressed as they cooled beyond the peak of metamorphism, a situation that would appear to require special tectonic conditions. This proposed "counterclockwise" P-T path for the Merrimack belt is in sharp contrast to the more traditional "clockwise" path (England and Thompson, 1984; Thompson and England, 1984) for the adjacent Bronson Hill anticlinorium (Figure 10). Tectonic correlations between the two regions would suggest that subsequent to tectonic loading in the nappe stage, rocks of the Bronson Hill anticlinorium were progressively unloaded, while those of the Merrimack belt were being progressively loaded, even in late stages where they were already undergoing cooling. Very recent K-Ar mineral ages on hornblendes from this region by T. Mark Harrison at SUNY Albany (pers. comm., 1987) also hint at early cooling in the east, and later uplift and cooling in the west, that may relate indirectly to these contrasting P-T trajectories.

## DETAILED INTERPRETATION OF MOUNT GRACE AREA

In the past two years Robinson and M.S. student George Springston have been studying the implications for the Mt. Grace area (Figures 4,5) of the new stratigraphic and structural concepts developed by P.J. Thompson (1985 and this volume) in the Monadnock area to the northeast and by Elbert (1985, 1987, and this volume) in the Bernardston-Hinsdale area to the northwest. In the western part of the Mt. Grace area (Figure 4) the inverted sequence of Silurian Clough Quartzite overlying Devonian Littleton Formation is again confirmed as the inverted limb of the Bernardston nappe. This can be directly connected to the inverted limb with fossiliferous strata at Bernardston, once about 5 km of displacement on the Mesozoic Connecticut Valley Border Fault is restored (Figure 3). Further, the doubled-over layers of Clough Quartzite at Bliss Hill (Figures 4, 5, 6) are confirmed as the synclinal hinge of the Bernardston nappe, here plunging south.

The large area of gray-weathering schists surrounding the Tully gneiss dome (Figure 4), that were once assigned to the Littleton Formation, are now mainly assigned to the Lower Silurian Rangeley Formation in direct correlation with the Monadnock area. These include a few isolated small areas of conglomerate previously unknown or previously assigned to the Clough Quartzite. Within this area north of the Tully dome Springston has been able to subdivide the Rangeley according to the members previously worked out by Thompson (Figure 5). These outline folds roughly parallel to the northern end of the dome, but with an uncertain relationship to it.

The west margin of the Rangeley Formation is marked by the Brennan Hill thrust as defined by Thompson. This traces from the Monadnock area southward into the Mt. Grace quadrangle where it undergoes a complex involution, principally caused by the northward overturning of the main and Tully bodies of Monson Gneiss, and then runs southward along the east margin of the main body toward Connecticut. Detailed work in the vicinity of the Brennan Hill thrust in the central part of the Mt. Grace area (Figures 4,6) has produced some surprising discoveries.

On the lower side of the Brennan Hill thrust the Bernardston nappe is mostly cut out by the thrust from Bliss Hill southwestward for about 5 kilometers (Figure 6). At different places in this distance the Brennan Hill thrust rests on inverted Clough Quartzite, on a thin strip of Partridge Formation in the core of the nappe, and locally on Littleton Formation beneath the nappe. Around the Tully dome, as presently interpreted, the Brennan Hill thrust cuts much deeper, resting on Ammonoosuc Volcanics or directly on the Monson Gneiss of the Tully dome.

On the upper side of the thrust, Rangeley Formation predominates, but locally other Silurian units are present and near the northern termination of the main body of Monson Gneiss, the thrust appears to cut

down into pre-Silurian strata (Figure 4). To the south and to the southwest of Bliss Hill (Figure 6), the younger Silurian units include several lenses assigned to the Perry Mountain Formation mainly because of their iron formation boudins, one lens of Francestown Formation, and two lenses of Warner Formation. All of these younger Silurian strata are localized close to the Brennan Hill thrust in such a way as to suggest that the Rangeley was structurally inverted above them, probably on the overturned limb of an earlier nappe-stage anticline that subsequently was followed by the thrust. Near Butterworth Ridge northwest of the Tully dome (Figure 6) there is a narrow complex belt including confused representatives of all the Silurian units including iron formation of the Perry Mountain. So far this zone has not been adequately worked out except in one area where a pace and compass map is complete (Figure 6B). This shows a very tight syncline with two members of the Warner Formation, Perry Mountain Formation with iron formation, Rangeley Formation, Partridge Formation, and Ammonoosuc Volcanics. Francestown Formation is missing. In this vicinity, the Brennan Hill thrust is apparently along the contact between Rangeley Formation and Partridge Formation. At the northwest corner of the Tully dome (Figure 6), more extensive Warner Formation appears between Perry Mountain Formation and Rangeley.

Near the north end of the main body of Monson Gneiss the Brennan Hill thrust appears to cut through the base of the Rangeley into augen gneiss member of the Partridge, and some miles south, into the sulfidic schist member (Figure 4). In this vicinity an earlier recumbent anticline with a core of Monson Gneiss, the North Orange nappe, is believed to be in the upper plate of the Brennan Hill thrust. An anticlinal hinge of this fold nappe is exposed in the southeast corner of the Mt. Grace quadrangle (Figure 4). The anticlinal nappe also appears in a continuous belt around the southwest and east sides of the Tully dome, with two moderately well exposed anticlinal hinges (Figures 4 and 6). The placement of these recumbent fold nappes in the upper plate of the Brennan Hill thrust appears to be necessitated by the configuration of the thrust southwest of Butterworth Ridge (Figure 6) where Rangeley Formation is resting directly on Monson Gneiss. However, there is a problem with this interpretation in that there are two lenses of Perry Mountain Formation along the northeast margin of the augen gneiss of the Partridge, seeming to support location of a thrust in this position instead.

To understand the outcrop pattern and structural evolution of the Mount Grace area it is first necessary to recognize that the backfold-stage and dome-stage deformations that involuted the earlier structural features were themselves complex. This is illustrated in the structural relief diagram of the entire Orange area in Figure 7A which shows the pattern and local overturning of even the simpler gneiss domes, as well as the northward overturning and longitudinal transport of the main and Tully bodies of Monson Gneiss. In Figure 7A the "swirl" in the pattern of lineations and minor folds is illustrated schematically along the east side of the Warwick dome and Kempfield anticline. In the core of the Pelham dome it can be proved that the lineation is parallel to the transport direction of a set of dome-stage minor folds (Onasch, 1973). One can take this as a cue and treat all of the dome-stage lineation as being parallel to the transport direction of the gneiss domes in the same manner as salt domes. This assumption yields the schematic grand pattern of transport directions illustrated in the restored structural relief diagram in Figure 8. However, this is misleading because the northward overturning and transport of the main and Tully bodies had to precede the main dome stage because strata inverted during this longitudinal transport were refolded by major anticlines and synclines of the main dome stage, such as the Williams Pond syncline.

The refolding of early southeast-directed recumbent folds, the Oak Hill recumbent syncline and the Tully Brook recumbent anticline, by the dome-stage anticlines and synclines, as well as by the Camp Warwick dome, is illustrated in a structural relief diagram in Figure 7B. The relative age of this recumbent folding is uncertain, but it is tentatively assigned to the stage of backfolding on circumstantial evidence, in part because there are peak metamorphic fabrics in some amphibolites that are parallel to the exposed synclinal hinges of this fold system. Further, similar southeast-directed recumbent folds in the Pelham dome and its cover are truncated by the Belchertown intrusion with a zircon age of 380 m.y. that was itself deformed and metamorphosed in the dome stage. It is tentatively suggested that the southeast-directed recumbent folding may have been slightly earlier than the overturning of the main and Tully bodies of Monson Gneiss.

With the above features in mind, the progressive development of the structure in the central part of the Mt. Grace quadrangle is illustrated in a series of tectonic cartoons (Figure 9). This shows 1) the formation of the Bernardston and North Orange fold nappes, 2) the thrusting of the North Orange nappe onto the Bernardston nappe along the Brennan Hill thrust, 3) the involution of the fold nappes and thrust nappes by

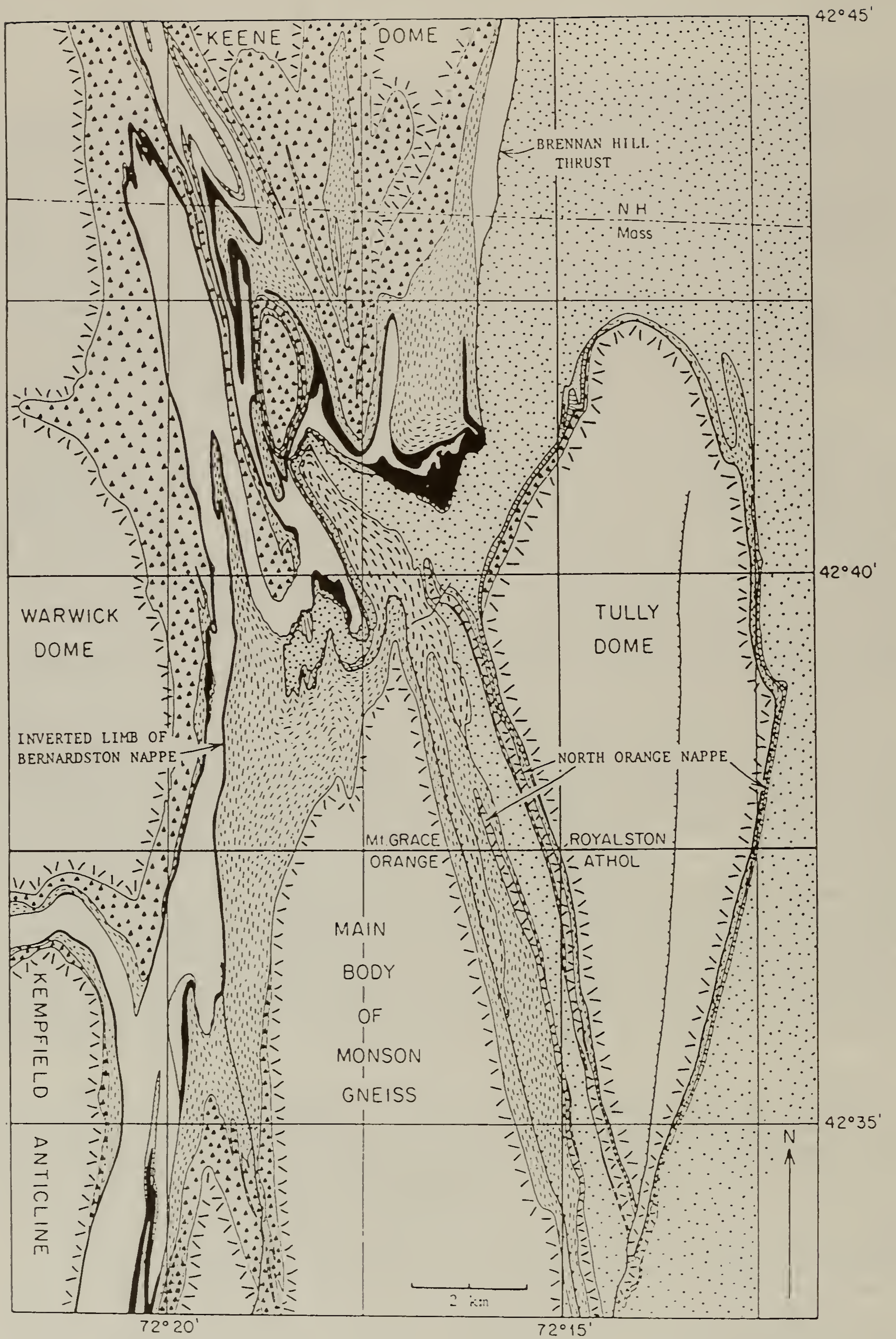



Figure 4. Generalized progress geologic map of the Mt. Grace quadrangle and adjacent parts of the Royalston, Orange and Athol quadrangles west-central Massachusetts and adjacent New Hampshire. For key to symbols see Figure 6A.

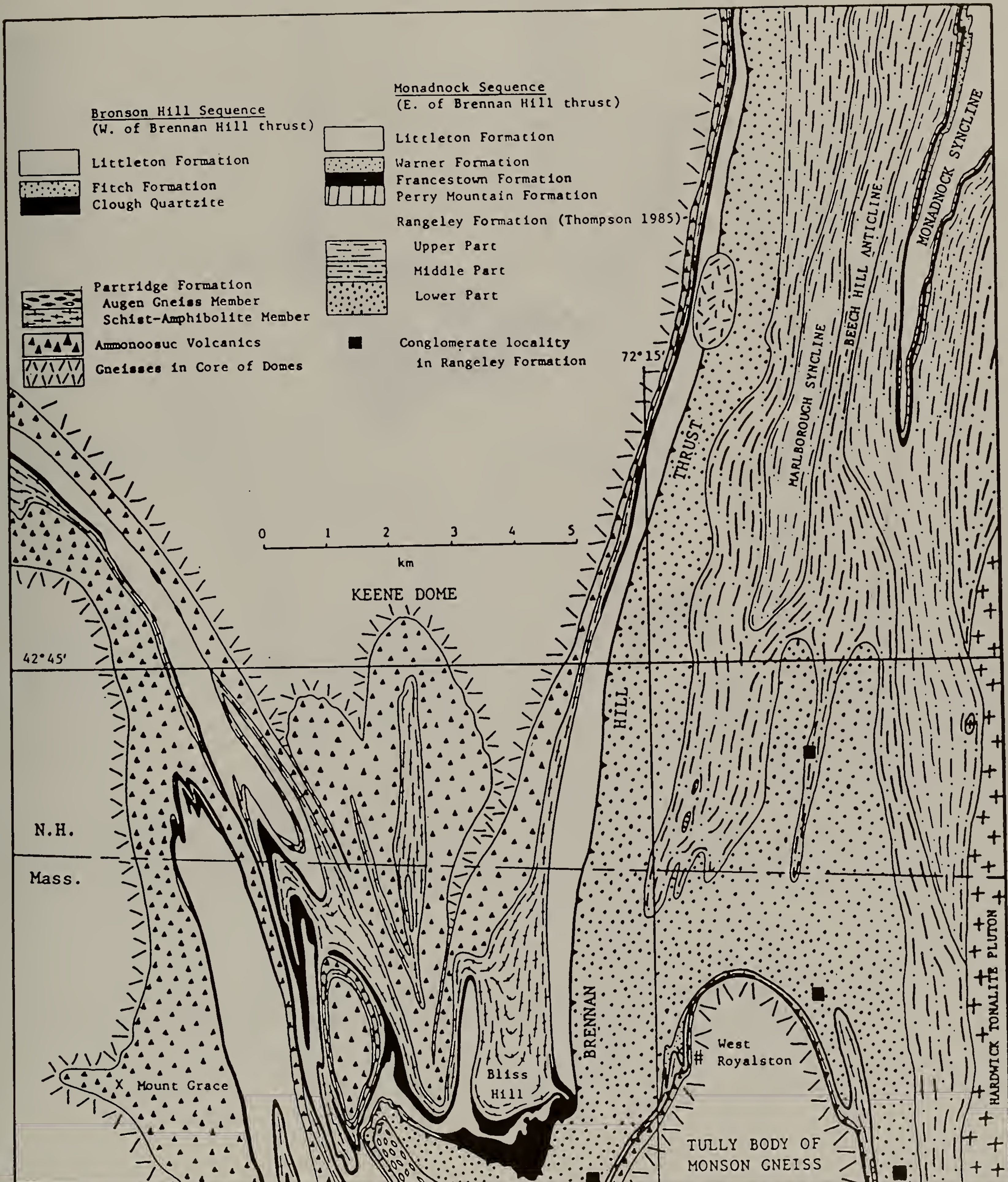



Figure 5. Distribution of members of the lower Silurian Rangeley Formation in the West Royalston area, Massachusetts - New Hampshire (George Springston, map in progress) and the adjacent Monadnock area (P.J. Thompson, 1985). Shows relationship to Brennan Hill thrust and Monadnock nappe-stage recumbent syncline.

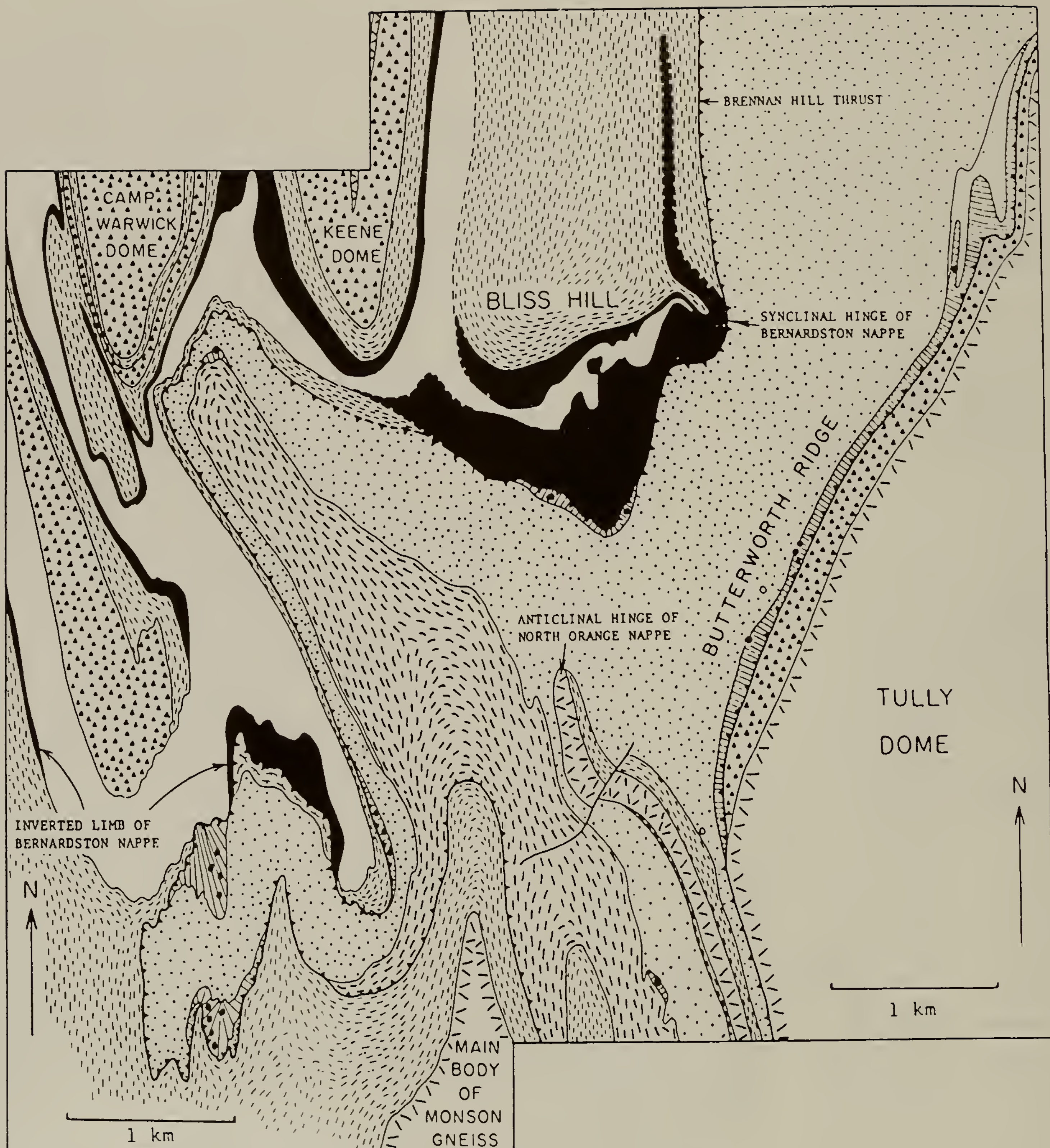



Figure 6. Detailed map of the central part of the Mount Grace area showing complex involution of the Bernardston nappe and the Brennan Hill thrust by northward transport of the main and Tully bodies ("Tully dome") of Monson Gneiss.

Figure 6A. Key to stratigraphic units in the central part of the Mt. Grace area.

| | AUTOCHTHON | BRENNAN HILL THRUST SHEET |
|---|---|---|
| LOWER DEVONIAN | LITTLETON FM. | |
| | FITCH FM | WARNER FM. |
| SILURIAN | | FRANCESTOWN FM. |
| | CLOUGH QUARTZITE | PERRY MOUNTAIN FM. WITH IRON FM. |
| | | RANGELEY FM. |
| MIDDLE ORDOVICIAN | PARTRIDGE FM. | PARTRIDGE FM. AUGEN GNEISS MEMBER SCHIST MEMBER |
| | AMMONOOSUC VOLCANICS | |
| ORDOVICIAN ? PRECAMBRIAN ? | GNEISSES IN CORES OF DOMES | MONSON GNEISS – NORTH ORANGE, CREAMERY HILL BANDS |

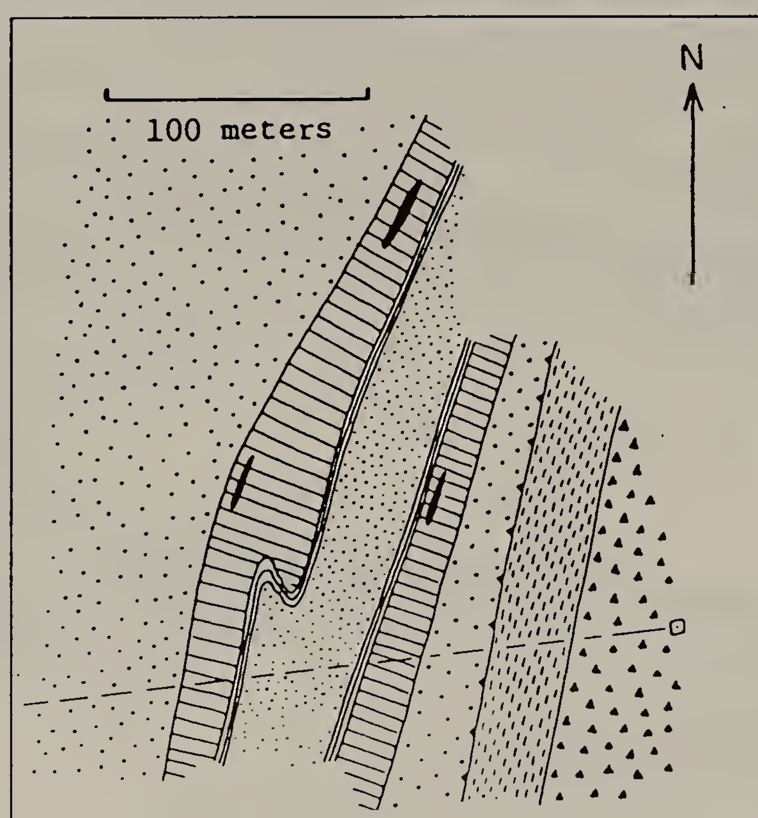
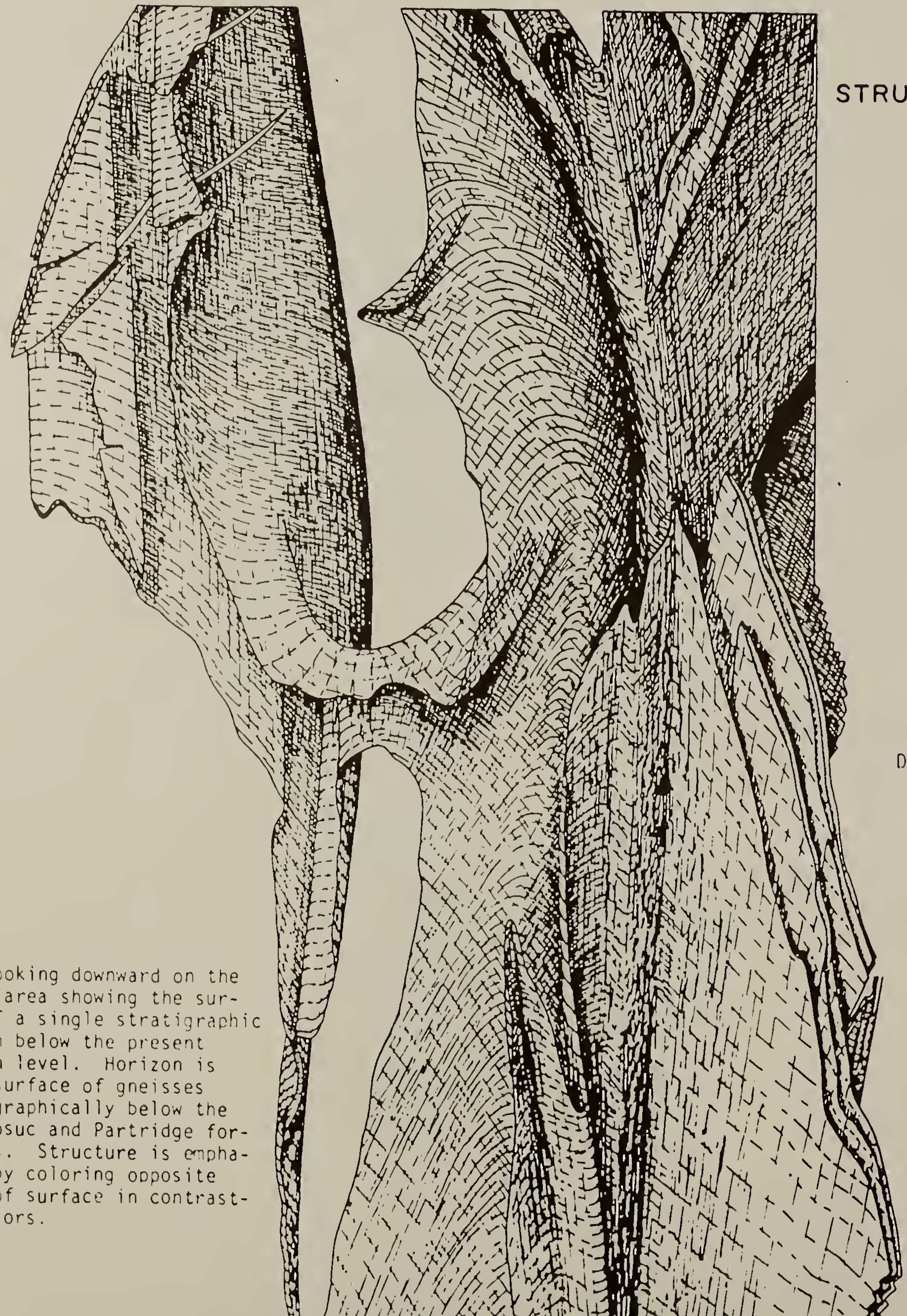



Figure 6B. Detailed pace-and-compass map showing setting of three boudins of Perry Mountain Formation on Butterworth Ridge (see Figure 6). Fine stipple shows upper biotite - feldspar granulite member of Warner Formation. Fine parallel lines indicate lower well bedded calc-silicate member. Francestown Formation is absent. Geology by Peter Robinson and George Springston.

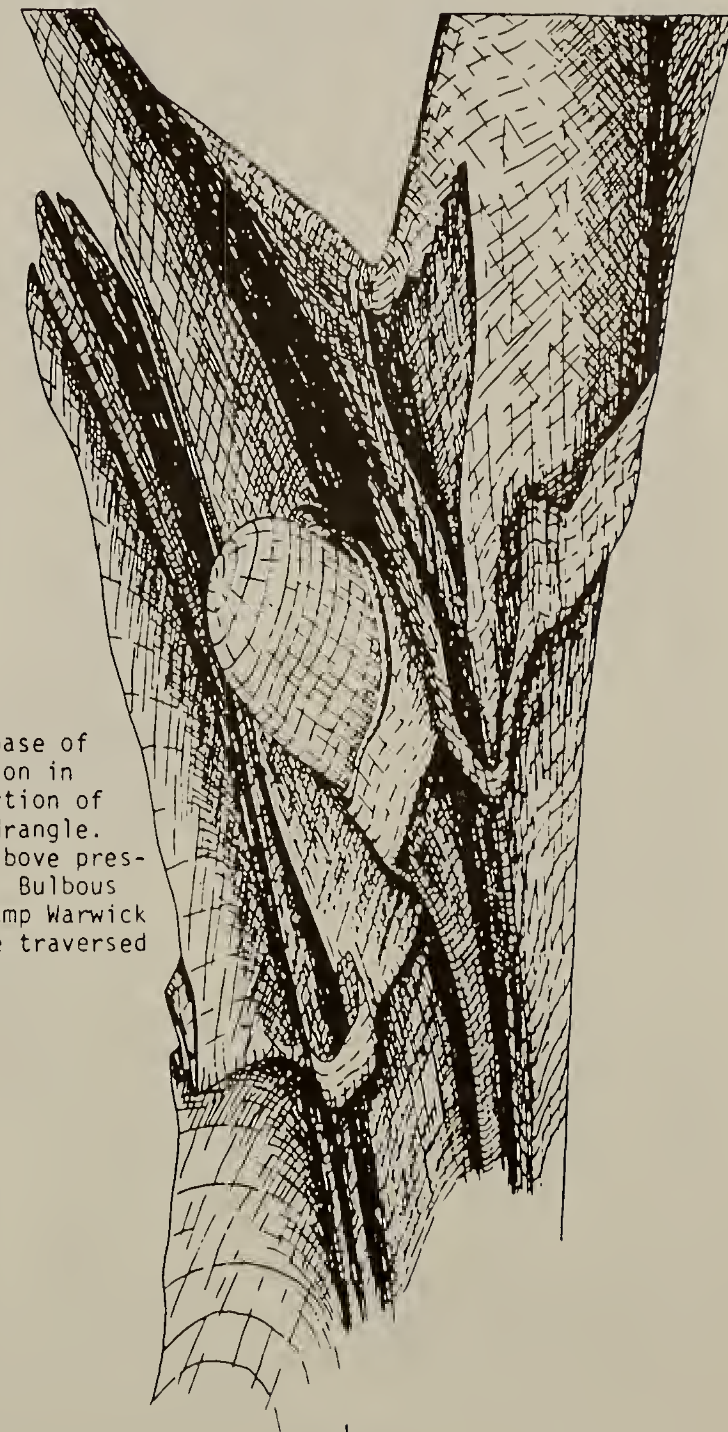
Figure 7

**STRUCTURAL RELIEF DIAGRAMS OF THE ORANGE AREA**

(from Robinson 1963)



View looking downward on the Orange area showing the surface of a single stratigraphic horizon below the present erosion level. Horizon is upper surface of gneisses stratigraphically below the Ammonoosuc and Partridge formations. Structure is emphasized by coloring opposite sides of surface in contrasting colors.



Detailed view of the base of the Partridge Formation in the north-central portion of the Mount Grace quadrangle. Partially restored above present erosion level. Bulbous structure is the Camp Warwick Dome which will be traversed on Stop 2.

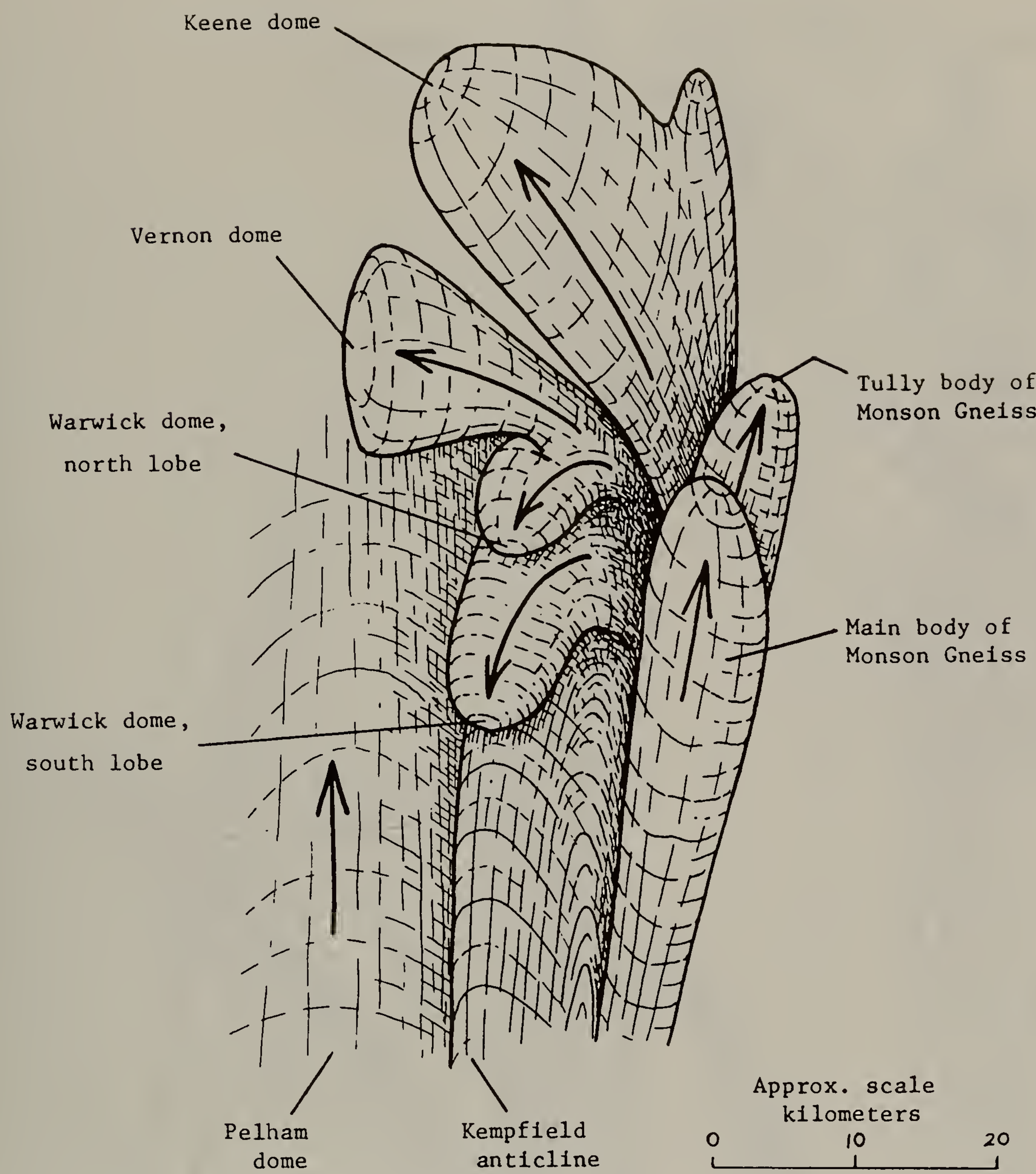
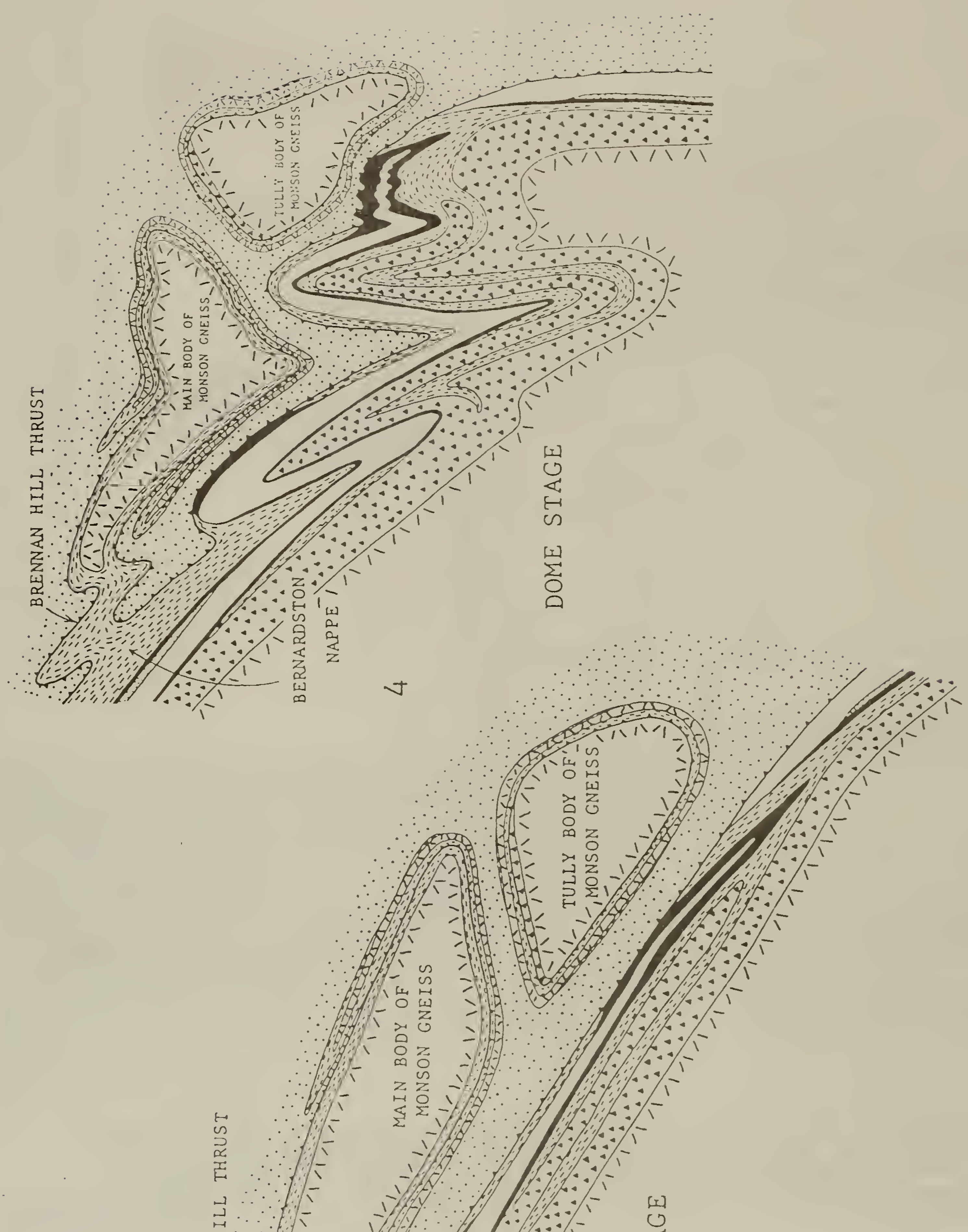



Figure 8. Schematic restored structural relief diagram showing inferred flow pattern of major gneiss bodies during dome stage. The "swirl" runs approximately N-S through the eastern part of the figure. Late asymmetric folds in the Pelham dome (Onasch, 1973) indicate dome cover was sliding southward, the same as the Warwick dome, hence relative northward motion of the Pelham dome core is inferred.


BRENNAN HILL THRUST
MAIN BODY OF MONSON GNEISS
TULLY BODY OF MONSON GNEISS
BERNARDSTON NAPPE
4
DOME STAGE
BRENNAN HILL THRUST
MAIN BODY OF MONSON GNEISS
TULLY BODY OF MONSON GNEISS
Oak Hill folds
3
BACKFOLD STAGE

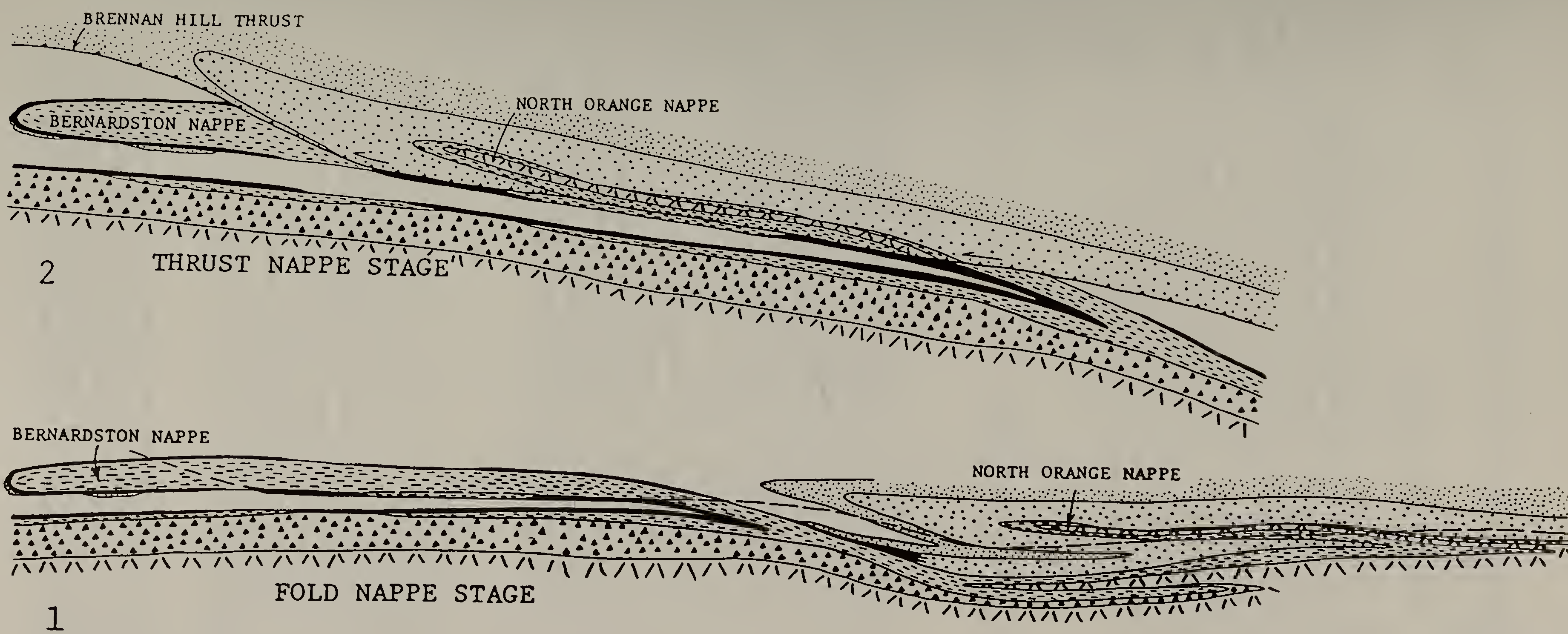



Figure 9. Schematic cross sections through the Mt. Grace area, Massachusetts, showing the sequence of structural development: 1) Formation of Bernardston and North Orange fold nappes. 2) Thrusting of North Orange nappe onto the Bernardston nappe along the Brennan Hill thrust. 3) Involution of fold and thrust nappes by northward transport and eastward overturning of the main and Tully bodies of Monson Gneiss. 4) Formation of gneiss domes and related tight folding. Actual timing of the east-directed Oak Hill recumbent folds in the Keene gneiss dome is unknown, but most recent indications are that these belong to the stage of backfolding.

northward transport and eastward overturning of the main and Tully bodies of Monson Gneiss, and 4) formation of gneiss domes and related tight folding.


## ACKNOWLEDGEMENTS

The first author wishes to acknowledge the support and interest of former students, professors and friends too numerous to mention here, many of whom have been acknowledged in greater detail previously. It is necessary particularly in the context of the interpretation in this paper to acknowledge the inspired and dedicated research of former stduents and students Peter J. Thompson, David C. Elbert, John C. Schumacher, Kurt T. Hollocher, Thomas M. Pike, Robert J. Tracy, Margaret Roll, and Robert D. Tucker. Marie Litterer worked over the years in preparing and lettering maps and figures. Completion of the present text and figures would not have been possible without the extensive last-minute selfless efforts of present students Henry N. Berry, Virginia Peterson, David C. Elbert, Jennifer Thompson, and Vincent DelloRusso. This research has been supported over the years by the National Science Foundation, Crustal Structure and Tectonics, and Petrogenesis Programs, most recently by Grant EAR-86-08762 (to Robinson). To each of these persons and institutions we give our grateful acknowledgements.


## ROAD LOG

Assemble at 8:30 A.M. at Kulick's Country Mall in Winchester, N. H. This lies on the left (north) side of Route 78 at 0.3 miles beyond the beginning of the road log.

<u>Mileage</u>

0.0 Road log begins at stoplights at junction of Routes 10, 119, and 78, which is 0.5 miles south on Routes 10 and 119 from the center of Winchester, N. H. For those proceeding from Keene the following distances and driving times are given: Begin junction of Routes 9 and 10 southwest of center of Keene. Proceed south on Route 10. Near Westport at 7.4 miles (11 minutes) on left is bush-up pavement outcrop showing folded biotite-rich mafic dike cutting previously deformed granite intruding gabbro, all part of the intrusive complex that composes the Swanzey gneiss. Center of Winchester and junction with Route 119 is at 12.1 miles (17 minutes). Proceed straight past lights in center on combined Routes 10 and 119 to stop lights atjunction with Route 78 (12.6 miles, 19 minutes from Keene). Turn left (east) on Route 7.

0.3 Assembly point , Kulick's Mall on left side of Route 78. Proceed east, then south, on Route78.

1.6 Broad view of topographic basin eroded in granitoid core of north lobe of Warwick dome. Folded cover sequence at north end of dome is exposed at Meetinghouse Hill in Winchestervillage and shows that north end of dome plunges due east. Mt. Grace (1617') looms to thesouth, held up by the Lower Member of the Ammonoosuc Volcanics in a major northeast-plunging cross fold separating the dome into north and south lobes. Reconstructions acrossthe Connecticut Valley border fault (Figure 8) suggest that the Vernon dome (see Elbert, thisvolume) is "rooted" in the northern part of the Warwick dome.

3.7 Massachusetts State Line.

5.8 Picnic grounds, Warwick State Forest, on right.

7.1 Bear left (east) off Route 78 at Warwick Public Library in center of Warwick and proceed easton main road. Northeast and north of the church are outcrops of the mafic Lower Member ofthe Ammonoosuc Volcanics cut by folded pegmatite dikes. The dominant rock type ishornblende amphibolite, but there are also epidote-rich layers, and plagioclase gneisses withcummingtonite, garnet, biotite, and secondary chlorite. These outcrops are typical of theLower Member of the Ammonoosuc Volcanics near Mt. Grace where the whole formation is4000 feet thick. In these outcrops minor folds and lineations plunge steeply northeast andthese outcrops lie west of the lineation "swirl"described above, whereas the outcrops atSTOP 1 lie east of it.

7.6 Park on right near small cut and outcrops of quartzite a few feet into woods. Small new roadcut in gray schist is visible ahead on both sides.

STOP 1A. See Figure 11. Briefly examine Clough Quartzite on east limb of Warwick dome.The purpose of STOP 1 in total (see Figure 11) is to demonstrate the Clough Quartzite in autochthonous position on the east limb of the Warwick dome, the Littleton Formation overlying it, the Clough Quartzite repeated in inverted position above the Littleton Formation in the inverted limb of the Bernardston nappe, and the Partridge Formation structurally above the Clough in the anticlinal core of the nappe. This doubled

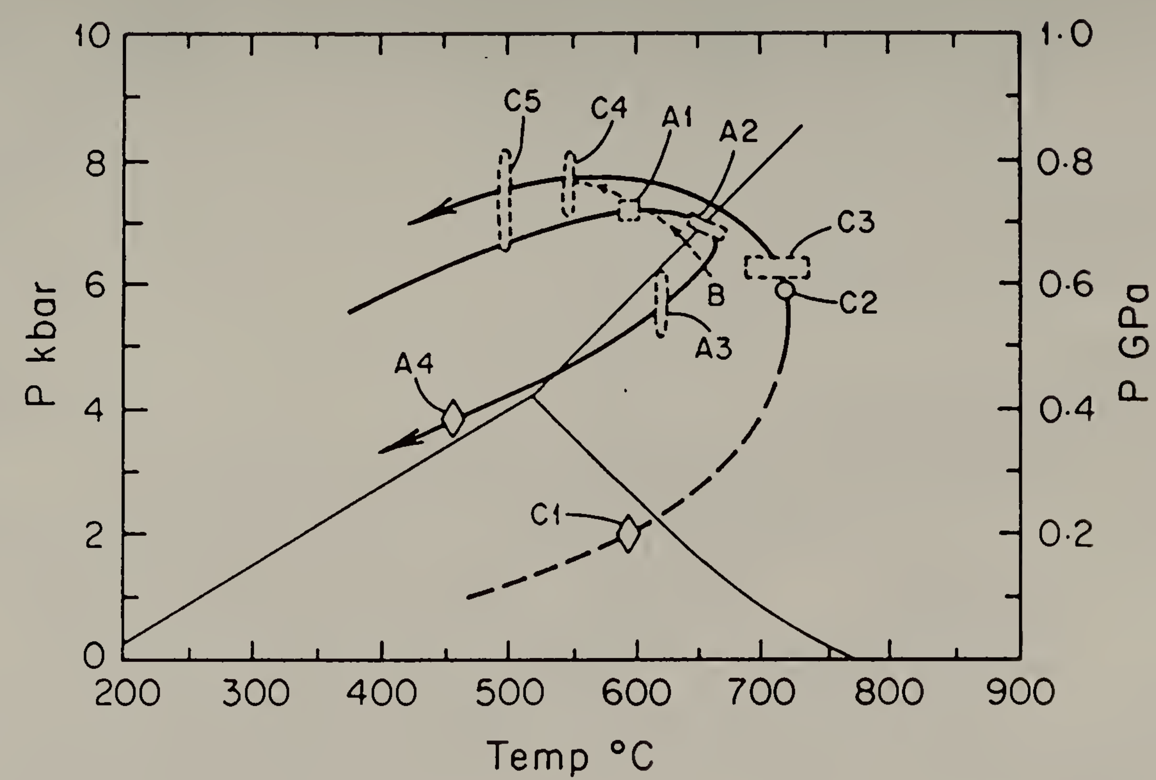



Figure 10. P-T trajectories from central Massachusetts and southwestern New Hampshire. A) central Bronson Hill anticlinorium near Keene dome based on the following: A1 - specimen I34I, garnet-biotite and garnet-cordierite thermometry, sillimanite-kyanite-quartz-garnet-cordierite barometry (Schumacher and Robinson, 1986, Table G-8); A2 -specimen 7AOBX, garnet-biotite and garnet-cordierite thermometry, kyanite-sillimanite-quartz-garnet-cordierite barometry (Schumacher and Robinson, 1986, Table G-8); A3 -cordierite enclaves, cordierite-spinel-quartz barometry with estimated temperature of 625°C (Schumacher and Seifert, 1986); A4 - kyanite-chlorite-quartz intergrowths in coarse cordierite (Robinson and Jaffe, 1969). B) eastern edge of Bronson Hill anticlinorium, Quabbin Reservoir area. Path of epidote formation, (Schumacher et al., 1988). C) Merrimack synclinorium, south-central Massachusetts based on the following: C1 - sillimanite pseudomorphs after andalusite, some of these occur in pegmatite indicating high T formation; C2 - beginning of breakdown of Fe-rich cordierite to quartz-sillimanite-garnet aggregate (data of Tracy and Dietsch, 1982); C3 - range of peak metamorphic T and P estimates from Zone VI (data of Tracy et al., 1976 and Robinson et al., 1982a); C4 - estimated P and T of mylonite recrystallization (Robinson et al., 1982a, 1986); C5 - estimated P and T of garnet rim re-equilibration in quartz-sillimanite-garnet aggregate inside zoned cordierite (data of Tracy and Dietsch, 1982).

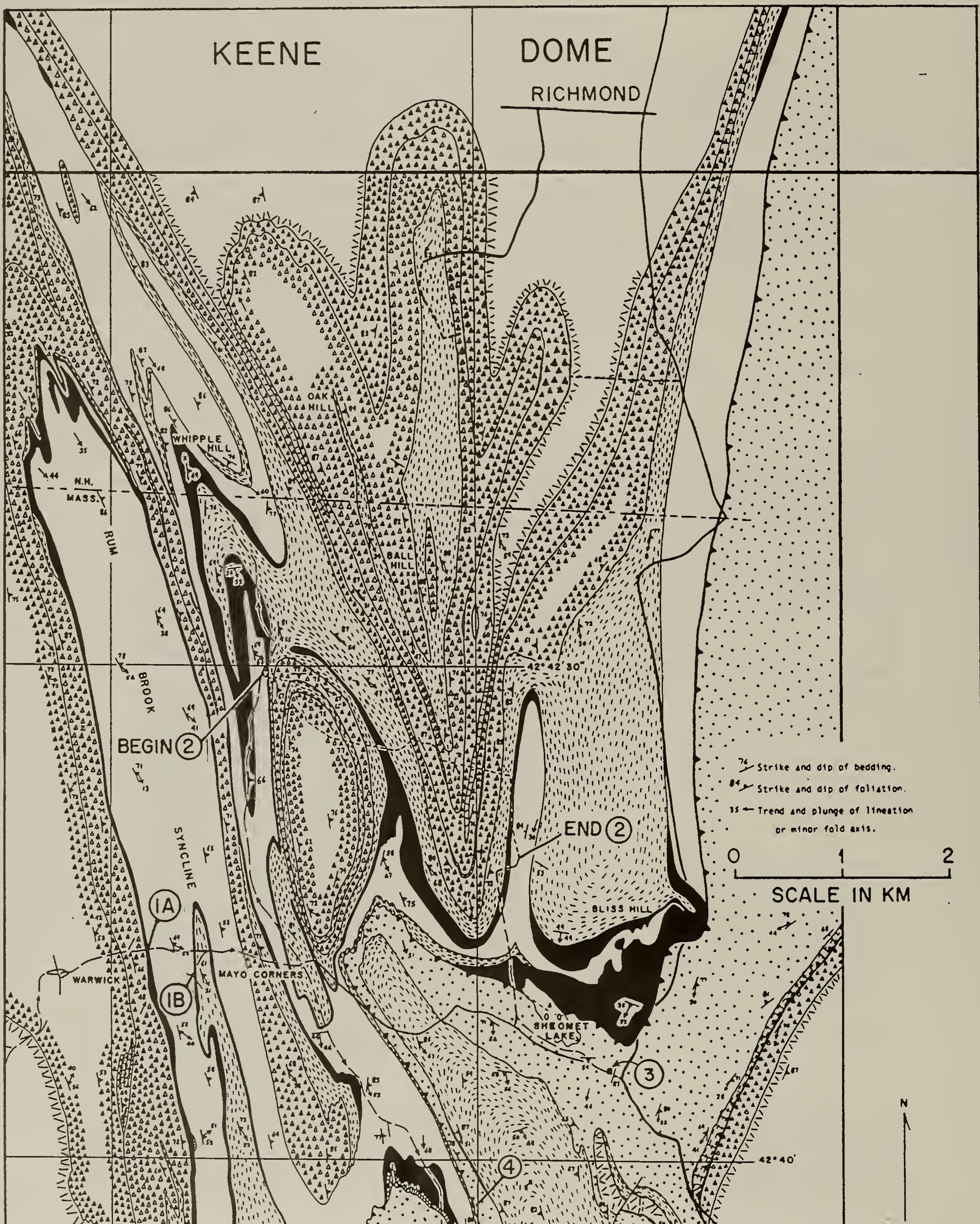
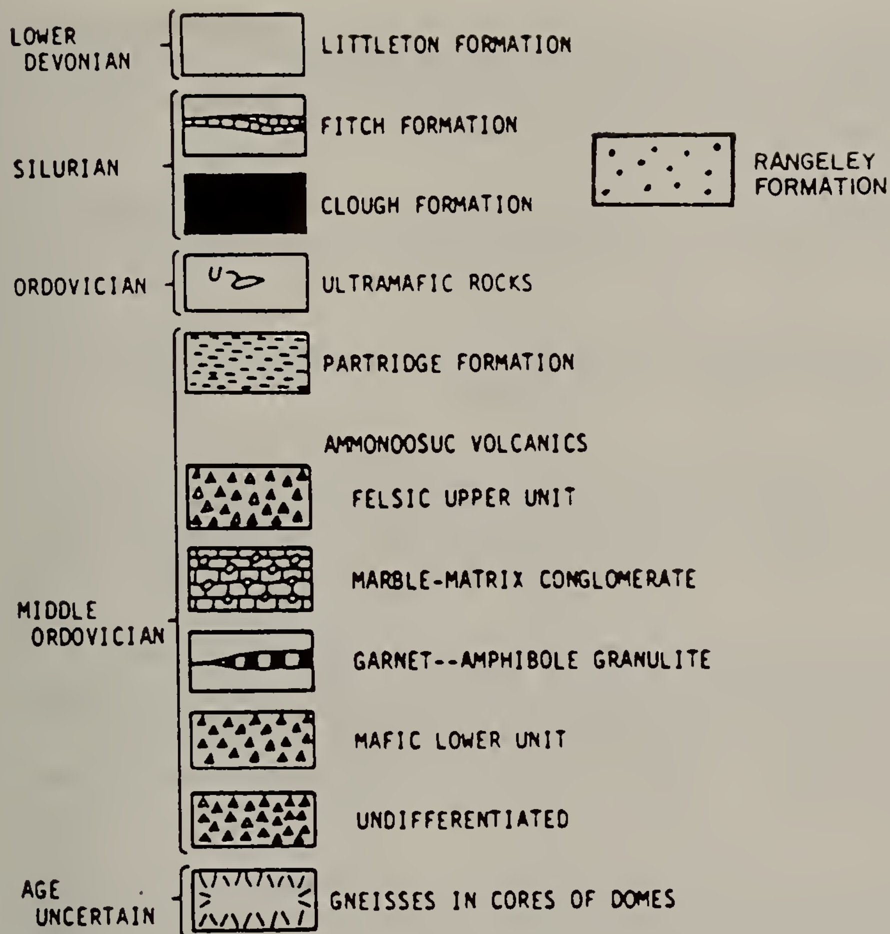



Figure 11. Detailed geologic map of the area of STOPS 1-4 in the Mount Grace quadrangle. Teeth are on the upper plate of the Brennan Hill thrust.

Figure 11 Legend.



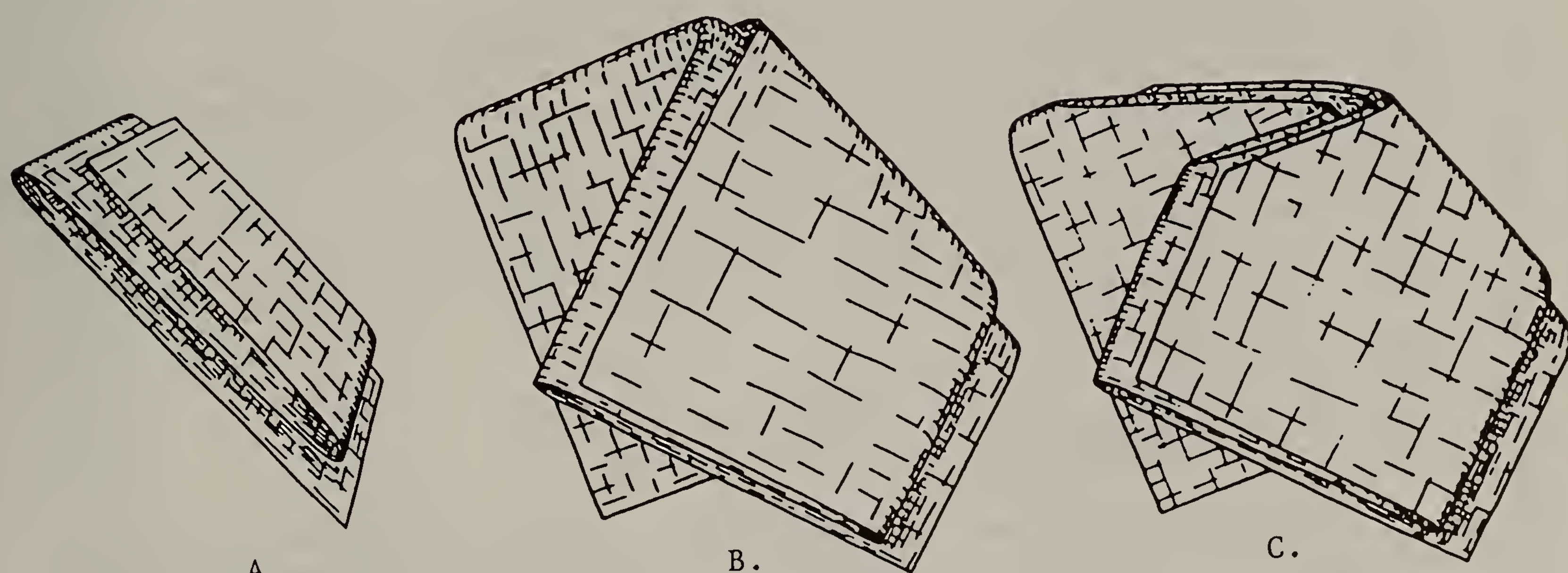




Figure 12. Origin of outcrop pattern at south end of Keene dome to be visited on STOP 2. Stratigraphic horizon shown is contact between Ammonoosuc Volcanics and Partridge Formation. A. Formation of Tully Brook recumbent anticline and Oak Hill recumbent syncline. Axial planes dip southeast and axes trend northeast. B. Folding of recumbent fold about Ball Hill anticline with axis plunging about 45$^{o}$ south-southeast. C. Erosion to give present outcrop pattern.

sequence produced by the nappe-stage recumbent syncline beneath the Bernardston fold nappe is isoclinally refolded by the dome-stage Rum Brook syncline. A cross section, in which movement on the Mesozoic Connecticut Valley border fault is restored (Figure 3), suggests that the inverted Clough described here connects directly in space with the lower limb of the nappe exposed at Bernardston.

Proceed east past small new road cut and larger road cut in Littleton Formation on both sides. These are good looking sillimanite-muscovite-staurolite schists that can be seen to greater advantage in natural exposures at STOP 1B.

8.0 Turn right (south) at intersection with Gale Road and proceed to partial road block at about 0.15 miles. Do U-turn and park on right side of Gale Road facing north.

STOP 1B. At this stop will be seen Littleton Formation on the west limb of the Rum Brook syncline, Partridge Formation of the Bernardston nappe in the center of the Rum Brook syncline, inverted Clough Quartzite on the east limb of the Rum Brook syncline, and finally Littleton Formation on the east limb of the Rum Brook syncline.

Walk south and west past road block in abandoned paved road to outcrops on north side of U-shaped bend. These are superb sillimanite-muscovite-staurolite schists of the Littleton typical of the lowest grade part of the sillimanite zone where quartz-sillimanite knots are just beginning to develop. Foliation dips 50-60° east. Dome-stage mineral lineation plunges steeply southeast in this location just on the east side of the dome-stage lineation swirl described above.

Return to position of parked vehicles, crossing unexposed inverted Clough Quartzite on west limb of Rum Brook syncline. Proceed east into woods just south of new house. Here there are several exposures of rusty-weathering sillimanite schist and amphibolite of the Partridge Formation in the center of the Rum Brook syncline. Proceed east through woods to steep east-facing ledges of Clough Quartzite including conglomerate overlooking swamp. Foliation dips about 60°. These are in the inverted limb of the nappe and on the east limb of the Rum Brook syncline. Follow Clough ledges north along strike and continue north to main E-W road. Just across road to northeast is a small outcrop of Littleton Formation on the east limb of the Rum Brook syncline. Walk short distance west to Gale Road, then south to vehicles. Drive north to main E-W road.

8.3 Turn right (east) off Gale Road onto E-W road and recross allochthonous Clough. Swamp ahead is underlain by Littleton, large outcrops (not visible) north of road.
8.5 Littleton-Ammonoosuc contact on east limb of Rum Brook syncline.
8.7 Small cut in sulfidic Partridge schist on left. This lies on east limb of dome-stage anticline of Ammonoosuc Volcanics. Between here and Mayo Corners is another narrow syncline of Littleton Formation with Clough on both sides, here separating autochthonous sequences peripheral to the Warwick dome to the west and the Keene dome to the northeast.
8.9 Mayo Corners. Turn left (north) on Richmond Road. View of Mount Grace to west. Leave as many cars as possible at this point to save pick-up time at end of Stop 2.
10.1 Camp Warwick on right; former CCC, now Forestry Work Camp. Clough Quartzite near nose of Whipple Hill digitation.
10.2 Pavement ends.
10.4 Cross powerline. View of reservoir on left.
10.7 Road runs over outcrop of Littleton schist on right.
10.8 Bottom of steep pitch, pavement begins.

STOP 2. This stop will take 2 to 3 hours and will cover a distance of approximately 2.3 miles, including an ascent of 270 feet, a descent of 480 feet, a second ascent of 170 feet, and final descent of 240 feet through open woods, wood roads, and a powerline. Boots are essential. One van will proceed to pick up drivers at end of stop and return them to the beginning.

Because the party will be strung out over a considerable part of the route at any one time, it is suggested that the route map (Fig. 13) and itinerary be followed closely. This is a wilderness area. "Drovers" will bring up the rear to make sure none are lost and try to assist any in distress. For those who have "had it" after station f or after station k an "escape route" that involves little further climbing can be

Figure 13. Detailed map of STOP 2.

pointed out. There is excellent collecting to be done on this traverse, particularly the later part, but everyone is URGENTLY REQUESTED TO REFRAIN FROM HAMMERING ON WEATHERED SURFACES THAT SHOW TEXTURAL AND STRUCTURAL DETAILS. Lineation and minor folds plunge very steeply southeast or south in most exposures and will not be described for each outcrop.

The details of STOP 2 are shown in Figures 11, 12, and 13. The major features of the outcrop pattern are similar to those shown by J. B. Hadley (1949), although the present stratigraphic and structural interpretation is far different. The present recumbent fold interpretation was first proposed by John Rodgers in conversations with Hadley and was independently deduced by Robinson from Hadley's map prior to field work in 1959, 1960, and 1961, when a wealth of confirmatory detail was turned up.

The essential feature of the present interpretation is a set of early recumbent folds in the Ammonoosuc and Partridge Formations, the Oak Hill recumbent syncline and the Tully Brook recumbent anticline. These have been refolded over three later structural features, the Camp Warwick dome, the Williams Pond syncline and the Ball Hill anticline. In the Ball Hill anticline, the Oak Hill recumbent syncline is outlined by a V-shaped area of Partridge Formation completely surrounded by Ammonoosuc. The northern ends of the "V' are interpreted as hinges of the recumbent syncline and show it originally had a northeast trend (Figure 12). The same recumbent syncline shows a completely closed ring of Partridge in Ammonoosuc where it has been wrapped over the Camp Warwick dome. In the Ball Hill anticline, the Tully Brook recumbent anticline has a core of Swanzey Gneiss of the Keene dome or the Lower Member of the Ammonoosuc Volcanics (Schumacher, 1988). Around the Camp Warwick dome the core of the recumbent anticline is composed of the Upper Member of the Ammonoosuc Volcanics.

The unique Camp Warwick dome is a flattened piston-like body with both ends plunging about 60° southeast (Figure 8). All lineations and minor fold axes are parallel to the piston axis, so that the body conforms in geometry to forms observed in salt domes (Balk, 1949), but similar steep lineations and minor folds are found in all structures in this part of the quadrangle. The movement sense of minor folds could be interpreted as the result of compressive flattening of a previously existing cylinder of rock.

The Williams Pond syncline is occupied by a narrow belt of Clough Quartzite folded on itself with a "thermometer bulb" termination to the north. The Clough is as little as 70 feet wide with a basal conglomerate resting on Partridge Formation on both sides.

Station a Proceed south along base of slope for about 200 feet to outcrops of Clough conglomerate holding up north-south rib. Long axes of pebbles plunge steeply east. Turn east across rib, cross gully beyond (Partridge Formation poorly exposed here), and begin ascent of main hill.

Station b Felsic Upper Member of the Ammonoosuc Volcanics with minor amphibolite in core of Tully Brook recumbent anticline. Layered peraluminous gneisses with garnet and muscovite are common. Minor asymmetric folds. Prominent flat jointing normal to lineation. Continue up and to right past several outcrops to wood road.

Station c First outcrop at high point in bed of wood road is Partridge Formation in inner ring (recumbent syncline) of Camp Warwick dome. Note movement sense indicated by minor folds. Proceed south just east of wood road along nearly continuous rusty Partridge outcrops. Contact with inner Ammonoosuc is poorly exposed when road rises again.

Station d Low knob outcrops west of wood road. Felsic gneiss and "porhyritic" amphibolite with a strong lineation of plagioclase patches. Continue south on wood road until powerline is barely visible, then bear diagonally left up hill to save elevation, and come out on powerline near hill top. Numerous outcrops of typical felsic Upper Member of the Ammonoosuc, commonly with brown-, red- or yellow-weathering character and containing sillimanite in some examples. Expansive view east and west from hill top. Begin descent to east along powerline.

Station e Contact of Ammonoosuc in core of Camp Warwick dome and inner ring of Partridge Formation. Bedding and lineation in Partridge is outlined by sillimanite knots. A thin section of the Partridge from here contains quartz 46%, andesine 20%, biotite 25%, muscovite 4%, garnet 2%, sillimanite 2%, graphite 1%, and minor sulfide. Proceed eastward into ring of felsic Upper Member of Ammonoosuc, including rare

outcrops of gray sillimanite-biotite schist, and then into outer ring of Partridge. Note movement sense indicated by minor folds in Partridge. One schist from this belt has the same minerals as the specimen described above plus about 25% orthoclase, another is similar but contains 20% muscovite and little or no feldspar.

Station f Basal cobble conglomerate of Clough Quartzite on west limb of Williams Pond syncline exposed a few feet east of Partridge Formation on south side of powerline. Not easy to see direction of long axes. Large outcrop to east is well bedded gray quartzose schist assigned to the Clough Quartzite in the axial region of the Williams Pond syncline, although it resembles some Littleton. Where exposures are more complete it appears to grade into conglomerate on both sides. Note pegmatite boudins involved with asymmetric folds. Proceed east past old foundation to north bank of small stream crossing powerline. Bear left (north) off powerline and follow north bank of stream.

Station g Partridge Formation in stream cascade under large trees, well east of Clough of Williams Pond syncline. Below flatter section of stream, felsic gneisses of the Upper Member of the Ammonoosuc Volcanics are exposed in a second cascade. Below this leave stream on south bank and continue downhill.

Station h Large blocks and some outcrop of mafic Lower Member of Ammonoosuc Volcanics including some beds of red-weathering gedrite gneiss. This lies in the axial region of the Tully Brook recumbent anticline on the east limb of the Ball Hill anticline. Follow flags east through evergreens across the west branch of Tully Brook.

Station i Small outcrops of well bedded garnet-grunerite quartzite that is characteristic of much of the Garnet-Amphibole Quartzite Member of the Ammonoosuc Volcanics and forms a distinctive marker unit between the Upper and Lower Members on both limbs of the Tully Brook recumbent anticline. Here it is on the lower limb as well as on the west limb of the Ball Hill anticline. Grunerite from this locality studied by Prof. Howard Jaffe has gamma=1.687 equivalent to FeO/(FeO+MgO) of 0.60. Go south along west-facing slope to powerline where there is a superb outcrop of finely laminated felsic volcanics of the Upper Member just east and structurally below the Quartzite Member. Turn left (east) on powerline to top of rise. You are now on the axial surface of the Ball Hill anticline.

Station j For those who wish to be convinced. Follow blazed trail north along trace of axial surface of Ball Hill anticline across inverted Upper Member of the Ammonoosuc Volcanics to outcrops of Partridge Formation that lie structurally beneath.

Station k Fine-grained quartz-microcline-oligoclase gneiss with white quartz-sillimanite nodules of the Upper Member of the Ammonoosuc. PLEASE DO NOT HAMMER WEATHERED OUTCROPS AND LOOSE BLOCKS ON SOUTH SIDE OF POWERLINE. This rock type is characteristic of the upper part of the Upper Member and can be traced entirely around the "V" of Partridge Formation. Similar rocks farther west in the Mount Grace quadrangle have kyanite and staurolite crystals in the cores of the nodules and have nodules 2 inches thick and 5 inches long. A thin section from this outcrop contains quartz 53%, microcline 35%, oligoclase 5%, biotite 2%, and minor muscovite and garnet. 5% sillimanite occurs only in nodules, but is seen in contact with microcline at nodule edges. X-ray measurements of muscovite laboriously separated from a rock like this from Ball Hill showed only 5% paragonite component (W. P. Freeborn, pers. comm. 1967). The nodules are tentatively believed to have originated as glassy clasts in a crystalline volcanic matrix that underwent hydrothermal alteration or weathering to quartz-kaolinite rock and was subsequently metamorphosed to quartz-sillimanite rock. Identical nodular rocks have been described at this same horizon in Connecticut (Lundgren, 1963, 1964) as well as in numerous localities in metamorphosed volcanics in Precambrian shield regions including the Adirondacks.

Proceed east across low ground where there are excellent low outcrops of amphibolites, including anthophyllite-bearing rocks. These belong to the Lower Member of the Ammonoosuc on the lower limb of the Tully Brook recumbent anticline, east of the axial surface of the Ball Hill anticline.

Station l Single low outcrop to right of tower on west-facing slope. Coarse-grained plagioclase gneiss of Swanzey Gneiss exposed near axial surface of Tully Brook recumbent anticline on east limb of Ball Hill anticline.

Station m Large outcrop on left at steepest place. Coarse-grained anthophyllite-garnet-biotite-hornblende gneiss with bed of hornblende amphibolite containing thin layer of diopside-labradorite-epidote gneiss. This is in the mafic Lower Member of the Ammonoosuc on the upper limb of the Tully Brook recumbent anticline. PLEASE LEAVE DIOPSIDE LAYER INTACT, SAMPLE ELSEWHERE IF YOU WISH. Anthophyllite gneiss contains quartz 15%, andesine 41%, anthophyllite 25%, hornblende 3%, cummingtonite trace, garnet 7%, biotite 7%, ilmenite 2%, and apatite trace. The diopside gneiss contains labradorite 42%, diopside 485, hornblende 5%, calcite 1%, epidote 3%, and sphene 1%. The epidote forms spectacular rims around plagioclase.

Above brow of hill and left of trail is small outcrop of Garnet-Amphibolite Quartzite Member between Lower and Upper Members on upper limb of Tully Brook recumbent anticline. Proceed over hill top to old farm road and turn sharp right (south) down hill. In bed of road are several exposures of schist of the Littleton Formation that have been eroded out since original mapping in the early 1960's. These are on the east limb of the Ball Hill anticline and indicate that the Clough Quartzite shown in Figures 11 and 13 should be moved slightly to the west.

Descend to pick-up point on Royalston Road. For those who arrive well ahead of the main group, two outcrops on opposite banks of Tully Brook are pointed out.

Station n Outcrops and blocks of Littleton Formation coarse sillimanite schist.

Station o Outcrops and blocks of Clough conglomerate on east limb of Ball Hill anticline.

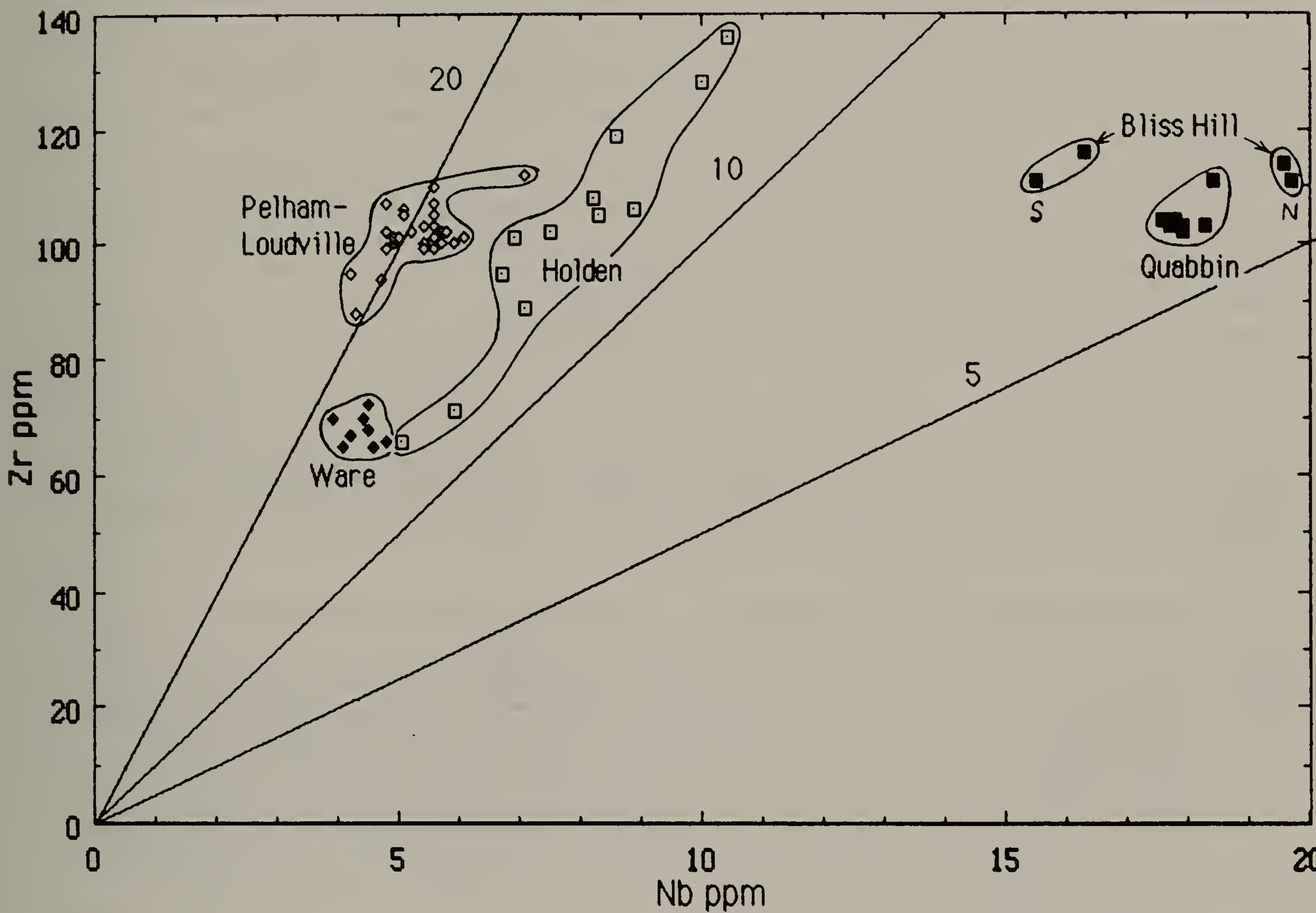
15.4 Road log commences at junction on Royalston Road (mileage to pick-up point included). Proceed west on Royalston Road past Stations n and o.
15.6 Turn left (south) at BM 258' on traveled road. Royalston Road deteriorates west of here.
16.5 Pavement begins.
16.6 Turn left through green gate at State Park entrance.
16.9 Cross brook outlet of Sheomet Lake. Park in lot on left near outlet or drive into lakeside parking. LUNCH STOP. Outcrops and blocks of Rangeley Formation sillimanite schist cut by south-dipping tourmaline veins east of lake outlet. East of the lake is Bliss Hill, underlain by complexly folded Clough Quartzite in a synclinal hinge of the Bernardston nappe. After lunch continue east through a second gate.
17.1 Turn left (east) on Athol Road.
17.2 Warwick-Orange Town Line. Park on right.

STOP 3. The purposes of this stop are two-fold, to examine the newly discovered south Bliss Hill tholeiitic diabase dike of early Cretaceous age, and to examine a large exposure of Rangeley Formation sillimanite-staurolite schist.

Walk west from the Town Line on the south side of Athol Road to exposure of dike in contact with rusty schist in recently built drainage ditch. The dike contains phenocrysts of Mg-rich orthopyroxene and plagioclase in a fine- to very fine-grained matrix of plagioclase, clinopyroxenes, and magnetite. Amygdules filled by calcite, other secondary minerals, and/or devitirified glass are characteristic. A primary flow structure is defined by oriented plagioclase microphenocrysts, and an internal chill has been found about 4cm from the southeast contact. Major- and trace-element analyses by XRF on two samples show that the rock is a quartz-normative tholeiite with the distinctively high Ni, Nb, and Sr, and "LREE-enriched" pattern on PMREE diagrams characteristic of all the Cretaceous intrusions in central Massachusetts (McEnroe, 1988) as compared to the Jurassic intrusions. On the basis of Zr/Nb ratios (Figure 14) the Cretaceous diabase intrusions must have had a source that was completely distinct from the sources of the more abundant Jurassic intrusions. These Cretaceous intrusions are the youngest igneous rocks in Massachusetts and the only known tholeiites in the Cretaceous province of New England and Quebec.

Walk east from Town Line on north side of road to large road exposure. This recently opened outcrop has been pronounced by Ben Harte of Edinburgh University as the most beautiful sillimanite-staurolite schist he has ever seen. Unfortunately, no electron probe analyses have yet been completed from this location although there is fairly abundant unpublished data from the surrounding region. The massive fibrolite sillimanite veins (described by B.K. Emerson, 1895, as "bucholzite") are here studded with euhedral staurolites.

Figure 14. Plot of Zr versus Nb in ppm with Zr/Nb ratios for Mesozoic diabases in central Massachusetts. The Holden and Pelham-Loudville systems are considered to be early Jurassic, the Ware system probably late Jurassic, and the Bliss Hill and Quabbin groups early Cretaceous.



The rock unit at this location was originally assigned by Hadley (1949) to the Lower Devonian Littleton Formation, then by Robinson (1963) to his Gray Member of the Middle Ordovician Partridge Formation, then assigned back to the Littleton Formation by Robinson (1967), and now on the basis of new work in the Monadnock quadrangle (P.J. Thompson, 1985) is assigned to the Lower Silurian Rangeley Formation! The steeply south-plunging open folds with parallel mineral lineation are typical of dome-stage folds near the south-plunging end of the Keene gneiss dome. One or two delicate graded beds suggest tops north indicating the strata are here upside down. This facing sense happens to be consistent with the appearance of lenses of iron formation of the Perry Mountain to the north, apparently in fault contact with Clough Quartzite across the Brennan Hill thrust.

This outcrop lies about one mile west of the staurolite-out isograd, although it has not yet been proved that the loss of staurolite is due to a prograde reaction or a change of bulk composition. Many schists in this vicinity contain only a muscovite-sillimanite-biotite-garnet assemblage. Probe analyses of staurolite assemblages in gray schists from this general vicinity give the following information:

| | | | |
|---|---|---|---|
| Staurolite | Fe/(Fe+Mg) = 0.83-0.84, | ZnO 0.27-0.54 wt %; | |
| Biotite | Fe/(Fe+Mg) = 0.53-0.59, | Ti/11 Ox. = 0.09-0.10; | |
| Muscovite | K/(K+Na) = 0.74-0.78, | Ti/11 Ox. = 0.01-0.02, | (Fe+Mg)/11 Ox. = 0.08-0.10. |
| Garnet Rims: | Alm 78-82, Pyr 10-13, | Spess 3-6, | Gros 2-5. |

A few garnets show growth zoning with decreasing spessartine from 13 in cores to 3 in rims, all at nearly constant pyrope content. Opaque minerals are uniformly ilmenite and graphite. Garnet-biotite geothermometry suggests, with considerable uncertainty, temperatures of 550-630 °C, and garnet compositions give estimates of minimum pressure of 5.4-6.5 kbar using the calibration of A.B. Thompson (1976b) and Tracy et al. (1976), respectively.

Do U-turn near schist outcrop and return west on Athol Road, bypassing park entrances.

19.3 Mayo Corners again. Turn sharp left (south) on Hastings Heights (hastingsites?) Road.

20.1 Outcrops on both sides at sharp bend in road. West side of road is formed of Clough Quartzite that is autochthonous on a tight anticline connected to the Warwick dome. Littleton Formation is exposed in a large knob east of road. On the Littleton outcrop, projecting quartz-sillimanite knots are elongated parallel to steeply plunging dome-stage folds in foliation. The rock consists of quartz 68%, biotite 15%, sillimanite 12%, muscovite 5%, minor garnet, ilmenite, and graphite, and a trace of zircon. Many Littleton schists in this vicinity contain staurolite as an additional phase. This schist is typical of the Littleton that lies in the syncline beneath the Bernardston nappe.

21.1 Small outcrops of Clough Quartzite on left. This is part of a thin remanent at this position of the inverted limb of the Bernardston nappe that is cut off immediately above by the Brennan Hill thrust.

21.2 Warwick-Orange Town Line.

21.5 Junction 985' with Gale Road. Bear left.

21.6 Sharp left turn off pavement onto Poor Farm Road.

21.8 Blue house on left and private road toward Johnsonian Pond. Depending on number of vehicles, walking tour for STOP 4 will begin here or if number of vehicles is small, it may be possible to drive part way in. Route beyond this point is described in walking log.

STOP 4. The purpose of this stop is to show some of the stratigraphic and structural features that led to the new interpretation of the Mount Grace area. Most of the detailed mapping was completed in 1966 and 1967, but stratigraphic reinterpretation is based on new data and interpretations in the Mt. Monadnock areas by P. J. Thompson (1985) and the Bernardston-Hinsdale area by D. C. Elbert (1985, 1986, 1987). The stop is designed as a traverse in a tectonically upward direction to see a series of features in sequence, all on the nearly vertical west limb of the dome-stage Williams Pond syncline. All lineations and minor fold axes, except at one outcrop, plunge 40-60° south-southeast parallel to the axis of the Williams Pond syncline. The one exception is the group of minor folds contained within a boudin of iron formation at station f that plunge 35-40° north and appear to belong to an earlier stage of folding. The features to be seen in sequence are:

1) Typical Littleton Formation from beneath the Bernardston nappe.
2) Sulfidic schist of the Partridge Formation with minor amphibolite boudins in a belt 100 feet wide that constitutes what remains of the anticlinal core of the Bernardston nappe. This is in direct contact with Littleton to the west, indicating absence or shearing out of Clough Quartzite in this location on the nappe.

3) Boudins of quartz pebble conglomerate with pitted matrix along contact between Partridge and gray feldspathic schist of Rangeley Formation to east. This is the inferred location of the Brennan Hill thrust. Along this contact to the south, large boudins of iron formation assigned to the Perry Mountain formation will be seen, indicating the stratigraphic sequence above the thrust is inverted.
4) A belt of gray feldspathic and sillimanitic schist and granulite assigned to the Rangeley Formation. This belt occupies an isoclinal syncline believed to have formed in the backfold stage during northward over folding of the main body of Monson Gneiss.
5) A narrow belt of feldspathic granulite-matrix conglomerate assigned to the lower member of the Rangeley Formation. This is in contact to the east with the Augen Gneiss Member of the Partridge Formation along a contact tentatively believed to be an unconformity .

From the blue house (see Figure 15) walk or drive north to fork in wood road. Where more travelled fork bears right bear to left (limited parking here). Continue on foot on left fork past old stone post marking town line between Orange and Warwick. Continue on wood road until evergreen trees end and continuous grass appears in middle of road. Turn right (north) across stone wall and walk about 200 feet north to large ridge-top outcrops.

Station a Gray homogeneous sillimanite-rich schist of the Littleton Formation. Quartz-sillimanite knots are evenly distributed and give a "braille" effect to the outcrops.

From these outcrops move diagonally down hill about S35E to east-facing outcrops overlooking small valley.

Station b This is a series of east-facing outcrops on west side of small valley to be traversed diagonally across strike in a direction N40E. Southernmost outcrop shows contact of gray Littleton schist to west and sulfidic Partridge schist of the Bernardston root zone to the east. At this location the root zone is 105 feet wide. The next several outcrops show sulfidic Partridge schist with small boudins of biotite amphibolite. About 200 feet to the northeast where the outcrop is steepest, sulfidic schist is in contact with a 1- to 2-foot lens of pebble conglomerate beyond which gray schist and granulite of the Rangeley Formation is exposed. A 10-inch pod of biotite amphibolite occurs in sulfidic schist about ten feet west of the conglomerate lens. The east edge of the sulfidic schist is the postulated location of the Brennan Hill thrust.

From base of highest outcrop walk south about 50 feet across shallow valley to north-facing outcrop.

Station c A thin layer of pitted conglomerate is in sharp contact with sulfidic schist to west and in contact with gray schist and feldspathic granulite to the east. This was originally interpreted as Clough Quartzite between Littleton and Partridge on the upper limb of the Bernardston nappe. However, the conglomerate and the gray schist and granulite are much more like rocks in the Rangeley Formation as defined in the Monadnock area. Traverse east across north slope and then northeast and downhill through a series of large outcrops of gray schist and feldspathic granulite of the Rangeley Formation, heading for a small southeast-trending stream gully shown on topographic map.

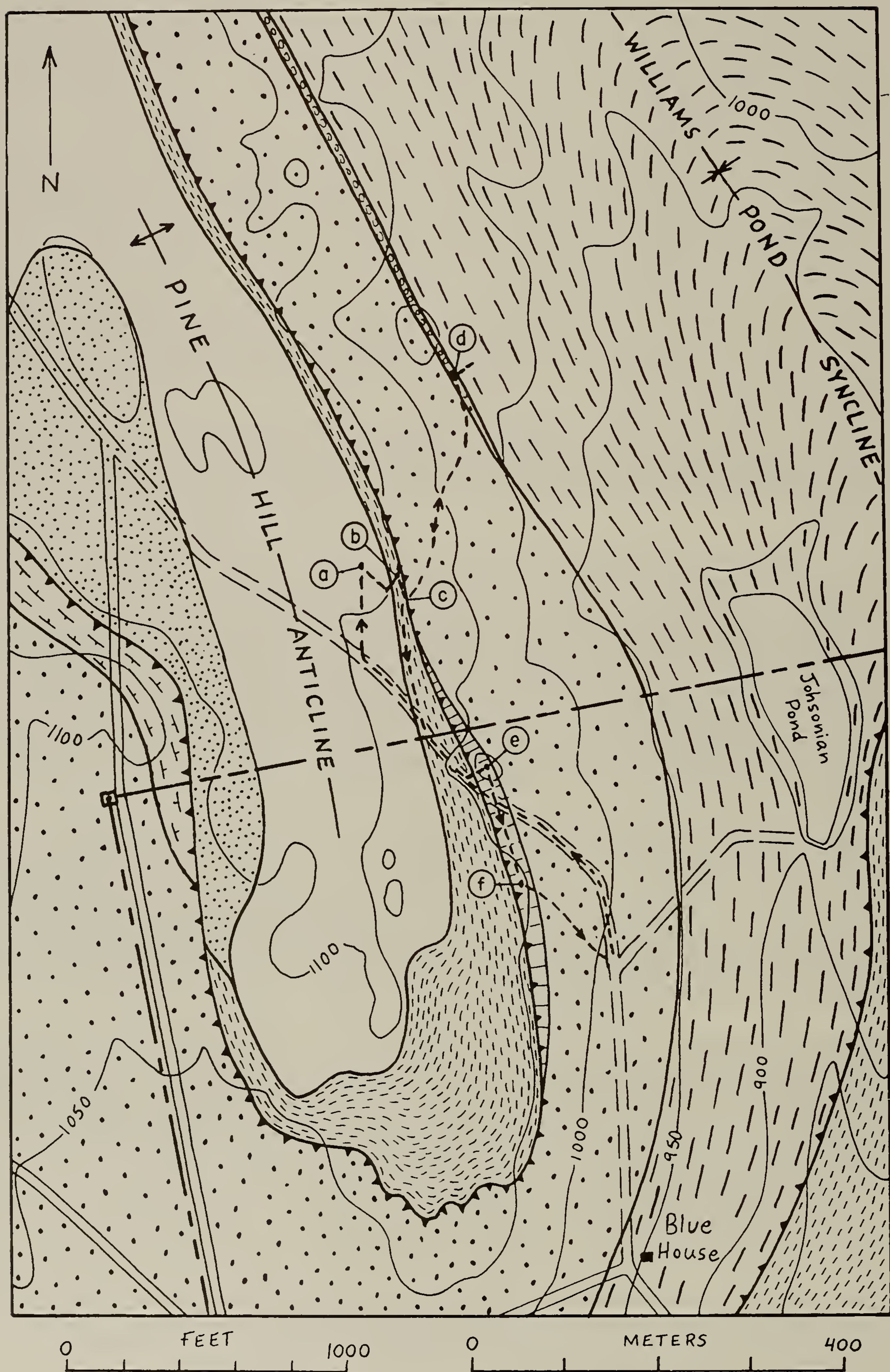
Station d The contact between gray Rangeley schist to the west, and rusty schist and augen gneiss of the Partridge to the east can be located just on the north side of the gully. From here for a distance of about 250 feet north there are small outcrops and numerous loose blocks of granulite-matrix conglomerate with a probable thickness of 20-40 feet. These resemble the granulite-matrix conglomerate describe in the the lower member of the Rangeley in the Monadnock area (Thompson, 1985) and here may be resting unconformably on the Partridge Formation. Walk east 100 to 200 feet to view typical exposures of the Augen Gneiss Member.

Retrace walking route to station c and follow shallow valley south to wood road. Proceed southeast on wood road past Warwick-Orange Town Line marker.

Station e Top of obvious small knob about 100 feet northeast of wood road. This shows the northernmost of several boudins of iron formation assigned to the Perry Mountain Formation based on the section at Biscuit Hill near Hinsdale, N. H. (see Elbert, this volume). It lies exactly along strike from the contact at station c.

Walk southward parallel to strike across wood road to small steep outcrop on south side. Here is exposed an unusual lens of conglomerate tentatively assigned to the Perry Mountain. It consists of dark

Figure 15. Detailed map of STOP 4 showing the Brennan Hill thrust above an attenuated anticlinal root zone of the Bernardston nappe.



pebbles in a quartz-rich matrix. The dark pebbles consist of garnet, grunerite, apatite, dark green ferric chlorite and a trace of magnetite. The matrix consists of quartz, apatite, grunerite, hornblende, and minor chlorite and garnet. The pebbles appear to be redeposited fragments from the iron formation. Continue south at same elevation across woods trail to high-standing knob held up by two large boudins of iron formation.

Station f This large outcrop was discovered by Robinson in 1966 and mapped in detail following appropriate cleaning by Huntington (1975). Figure 16A, based on his Figure 2 shows the distribution of rock types in the outcrop, including gently plunging early folds and the steeply plunging boudin neck line separating the two parts of the outcrop. Figure 16B shows details of delicately folded apatite-rich beds in the southern part of the outcrop. Except for the fact that the rocks are now assigned to the Perry Mountain Formation rather than the Littleton, there is little to add to Huntington's detailed mineralogical and petrological analysis. Yet to be accomplished is an analysis of the sedimentary environment and paleogeography at the time of deposition of these unusual rocks during the late early Silurian.

The low area west of the knob contains small outcrops of sulfidic schist and amphibolite of the Partridge Formation in the Bernardston root zone.

From the knob walk east on woods trail over pavement outcrops of feldspathic granulite of Rangeley Formation. Trail connects with fork in road. From there return south (right) back to blue house on Poor Farm Road.

Following Stop 4 continue driving down Poor Farm Road. Bottom of valley coincides almost exactly with position of synclinal keel of main body of Monson Gneiss.

22.7 Farm views on right showing valley eroded in main body of Monson Gneiss.

23.2 Junction and beginning of pavement at Athol Road. Turn sharp right (northwest) on Athol Road and descend again into basin of Monson Gneiss.

23.6 Turn sharp left (south) at bottom of hill onto North Main Street, Orange. On left is Williams Pond, for which the dome-stage structural syncline is named, here expressed as a south-plunging keel of inverted Monson Gneiss. Proceed south through basin eroded in Monson Gneiss. As one approaches northern outskirts of Orange there are fine views to the southwest showing the prominent topographic ridge of Partridge Formation and Ammonoosuc Volcanics that bounds the Monson Gneiss to the west.

27.9 Stop lights in center of Orange. Proceed straight through and cross bridge over Millers River.

31.4 Turn right onto eastbound entrance ramp of Route 2. Part way up ramp on left is cut in Monson Gneiss from which sample was collected on which R. E. Zartman obtained a zircon age of 440-450 m.y. Proceed east on Route 2.

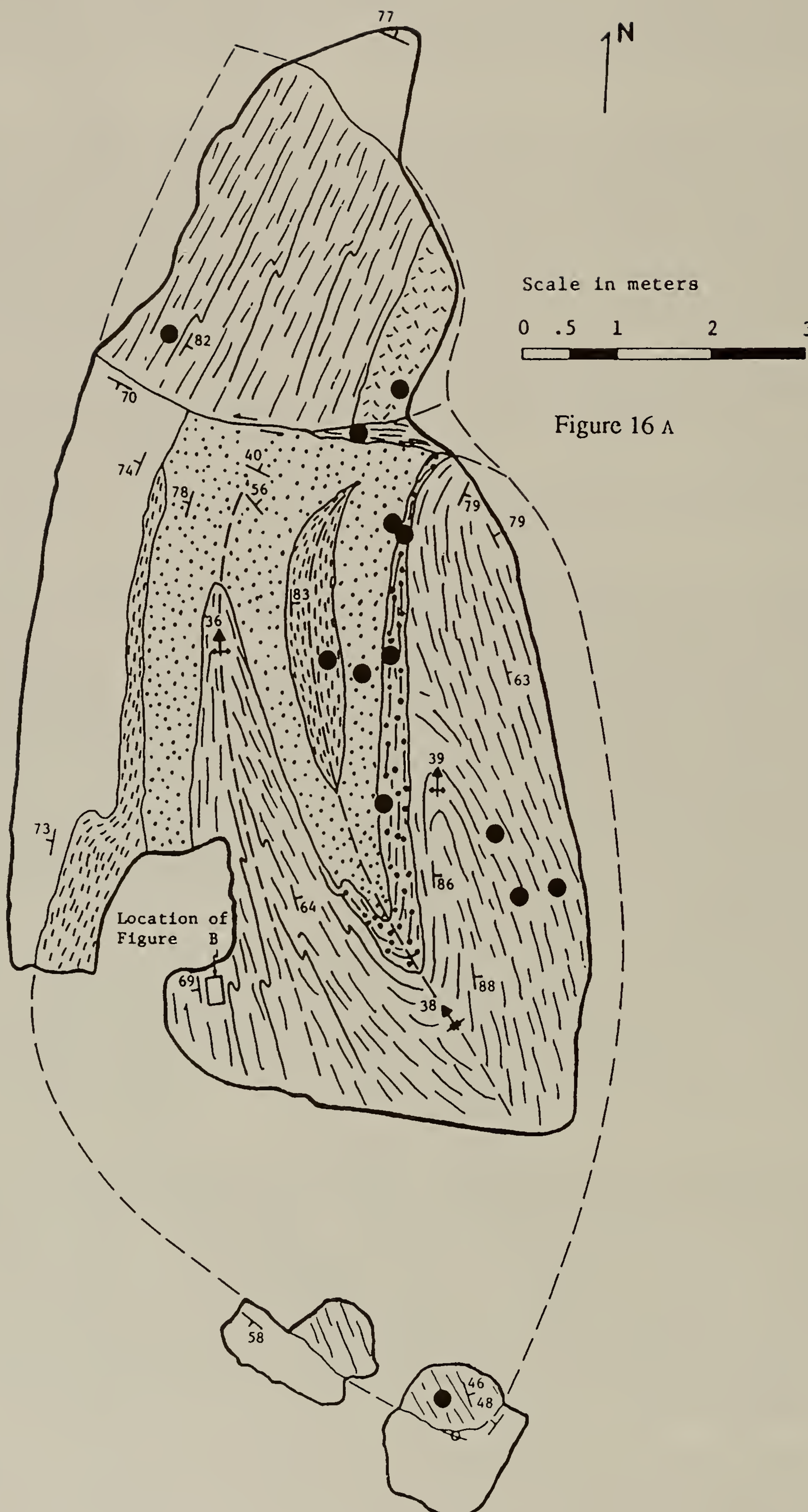
32.6 Exit for Route 202. Continue east on Route 2.

34.3 Bushed-over road cut on right. Gneisses in eastern part of main body of Monson Gneiss.

35.4 Athol road cut. Larger, long trench cut, both sides. STOP 5. If traffic permits, particularly if you intend to proceed west after this stop, bear left across highway to broad and firm grass strip on north side. This area is much safer and better for viewing the geology than the narrow area on the south side. Since Figure 18 was drafted in 1979 the south side of the Athol cut has been blasted back about ten feet. This has changed some details, but has not altered any essential features.

STOP 5. See Figures 17 and 18. This outcrop is dominated by a Mesozoic normal fault cutting the south end of the Tully body of Monson Gneiss. The south end of the Tully body is a simple anticline overturned to the east and plunging gently south-southwest. Dome-stage normal asymmetric folds occur on both limbs. The exposures at Stop 5 are the southernmost ones of the gneiss in the core of the Tully body. At this point the anticline is cut just west of its crest by a west-dipping normal fault, the Athol fault, bringing gray schists of the Rangeley Formation on the west limb down about 1300 feet into contact with Monson Gneiss of the core. The actual fault zone is about ten feet wide and contains gouge zones, hematite-stained and cemented breccia, intense silicification, and vuggy quartz veins, all features characteristic of known Mesozoic faults. Secondary alteration extends into the metamorphic rocks on both sides for a considerable distance. Here the fault strikes N4W, but regionally it strikes N15E.

On the east limb of the anticline, east of the gneiss of the core, is a thin layer of rusty Partridge schist which has been traced for many miles. Well bedded amphibolites at both contacts probably belong to the Partridge, although it is tempting to consider that they might represent the Ammonoosuc Volcanics highly attenuated. Next east of the schist and amphibolite is a layer of Monson Gneiss (Creamery Hill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38
58
46
48

Figure 16 A

Grunerite-garnet-pyrrhotite granulite

Olivine-magnetite granulite

Olivine-pyroxmangite granulite

Garnet-grunerite-quartz granulite

Mica-garnet schist

Sillimanite-muscovite schist

Sample location

Strike and dip of foliation

Trend and plunge of fold axis

Trace of axial surface of syncline

Trace of axial surface of anticline

Offset sense of fault in boudin neck zone

Fault

Known contact

Inferred contact

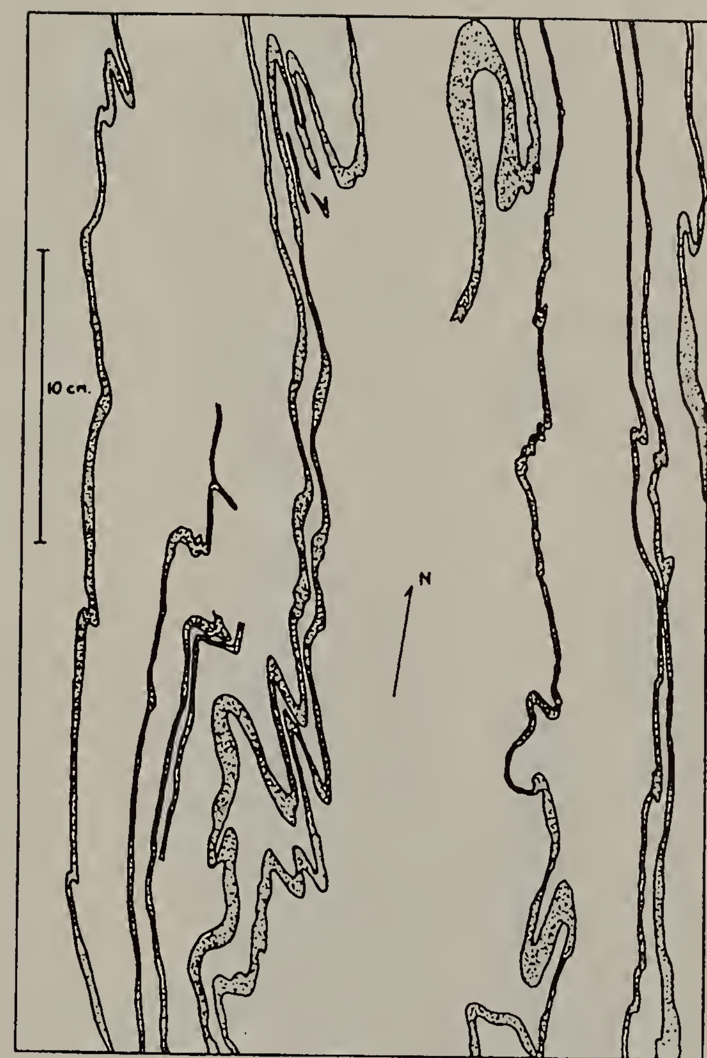



Figure B

Figure 16. Detailed outcrop map of boudins of iron formation in Perry Mountain Formation at STOP 4, station f (from Huntington, 1975). A. Outcrops of two boudins surrounded by sillimanite schist. B. Detail map of inset area showing complexly folded apatite beds.

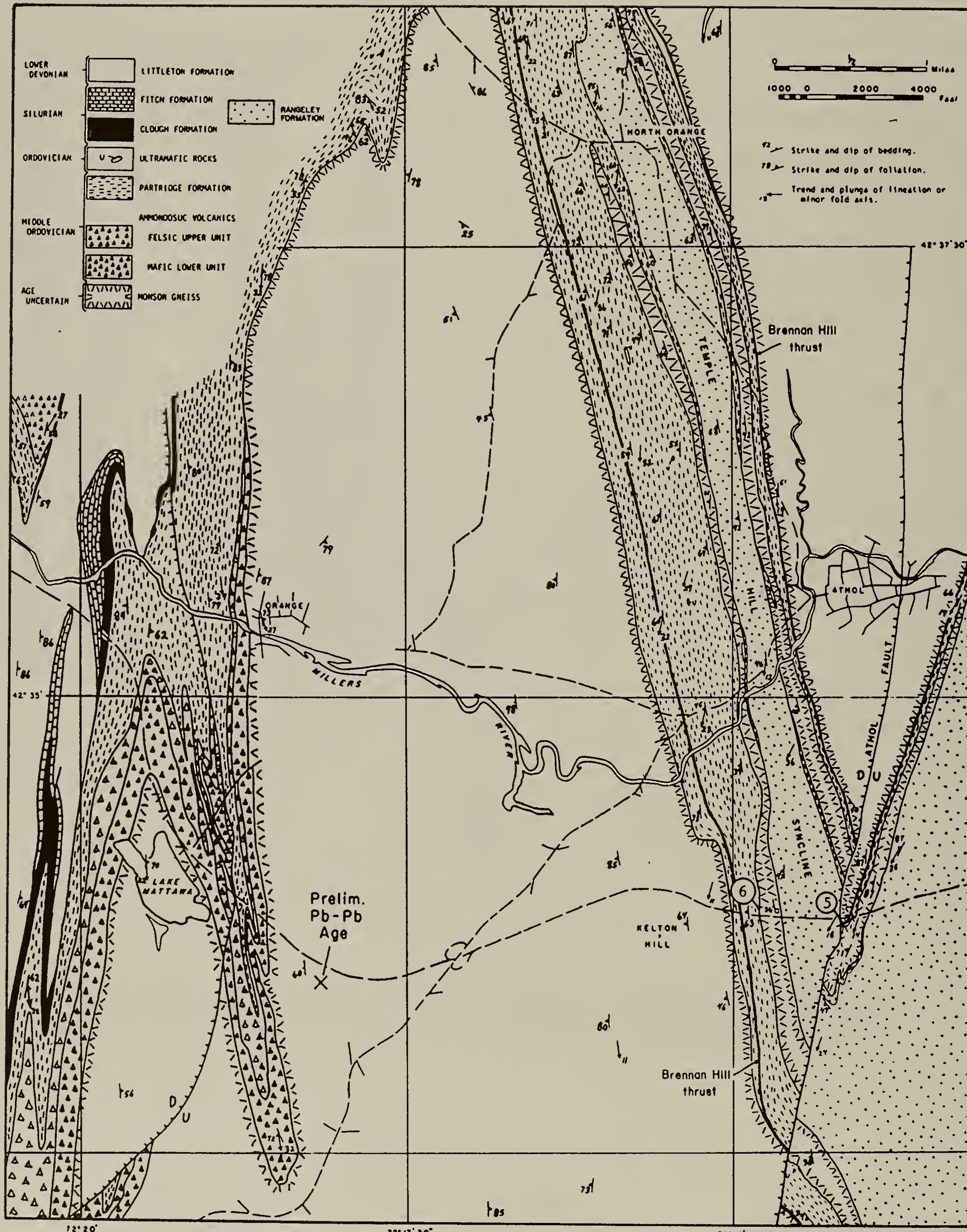
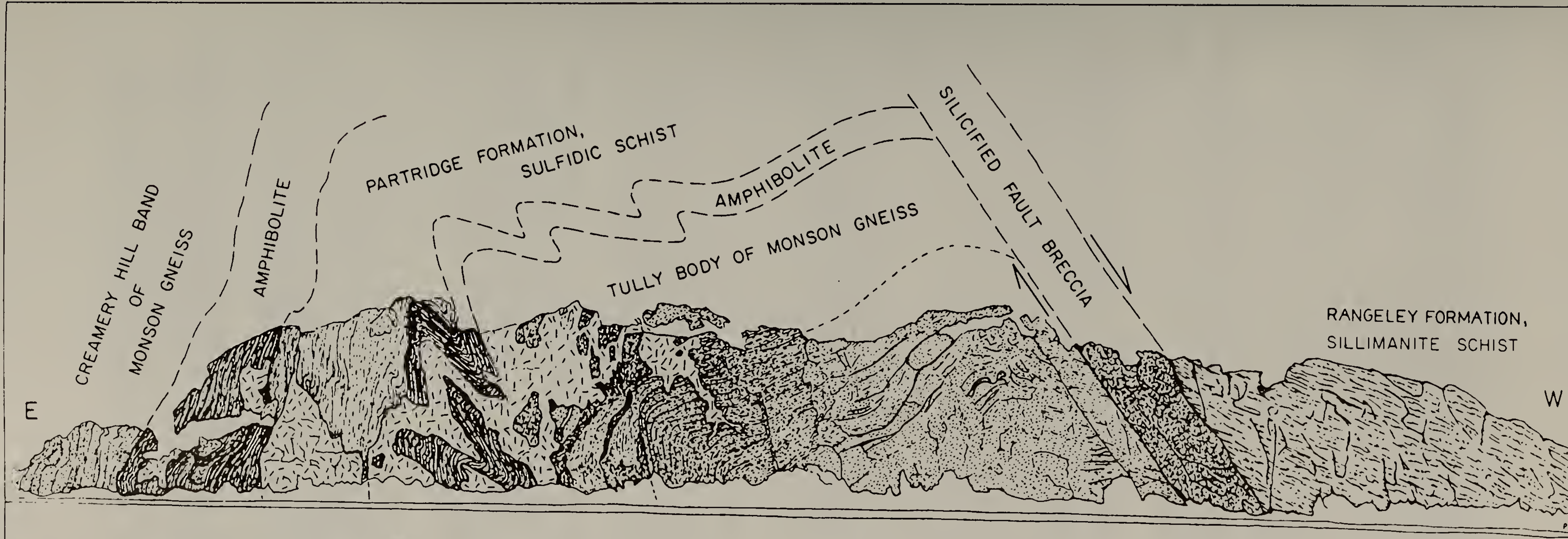



Figure 17. Geology of the eastern part of the Orange area, showing postulated position of North Orange nappe and Brennan Hill thrust in relation to STOPS 5 and 6.



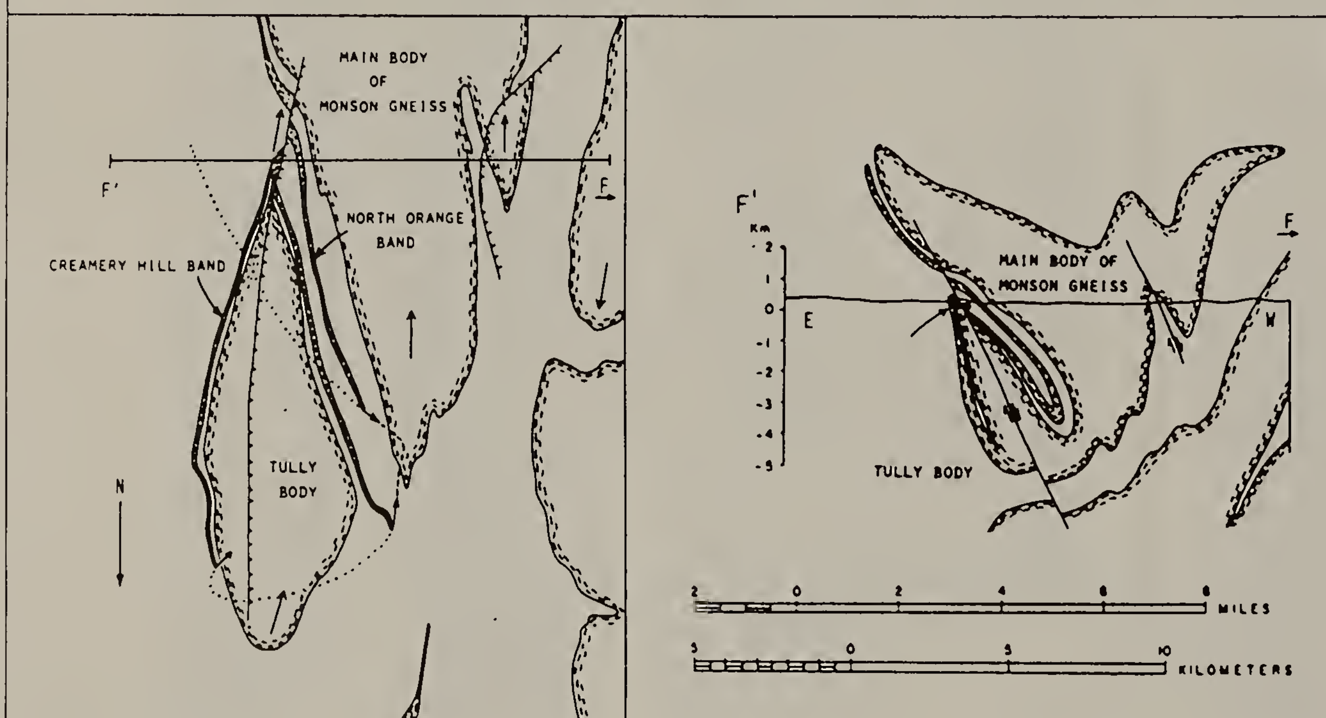




Figure 18. Sketch of south wall of rock cut at STOP 5. Mesozoic Athol fault cutting crest of Tully body of Monson Gneiss. Approximate location of cut is shown by tiny stippled rectangle in inset cross section. The major anticlinal or dome axis and satellitic folds and mineral lineations are parallel to a strong Beta maximum (10% per 1% area) with trend N22E, plunge 18SW (Robinson, 1963). The Creamery Hill band of Monson Gneiss in interpreted (see insets) as an extremely attenuated basement nappe separated from the main and Tully bodies by an extremely attenuated isoclinal syncline and by the Brennan Hill thrust. This can be interpreted as a fold of the same generation as the Bernardston nappe but lying tectonically higher and more easterly (see Figure 9). Drawing does not have constant scale. Outcrop is approximately 50 feet (15 meters) high at highest point.

band) that has been traced entirely around the southern part of the Tully body (see Figure 18) and is interpreted as an early anticlinal fold nappe of gneiss that formed contemporaneously with the Bernardston nappe. A similar band of Monson Gneiss (North Orange band) east of the main body of Monson Gneiss is interpreted as the same recumbent anticline repeated by folding about the Temple Hill syncline that separates the two bodies. The North Orange band has been traced from North Orange at least as far south as the northern edge of the Palmer quadrangle in southern Massachusetts. Three anticlinal hinges for this early fold nappe, now named the North Orange nappe, can be seen in Figure 17. This early fold nappe has been severely involuted, both by dome-stage folding and by the complex backfolding that included northward transport of the main and Tully bodies and the formation of the Temple Hill syncline itself (see Figure 9). Nevertheless a rational unwinding of these later deformations leads to the conclusion that the original transport direction for this nappe was from east to west. Further, if one accepts the tentative conclusion that the North Orange nappe lies in rocks structurally above the Brennan Hill thrust, then the trace of the Brennan Hill thrust must also lie in this outcrop, probably at or close to the contact between the narrow belt of Partridge Formation and the gneiss in the core of the Tully body.

The next outcrops to the east on Route 2 (visible from here) contain both gray and rusty schists of the Rangeley Formation on the east limb of the dome-stage anticline. The gray schists are similar to those on the west limb. These rocks are typical of a wide range of rocks above the Brennan Hill thrust. Where the schist has not suffered secondary alteration it consists of quartz 30-50%, biotite 20-40%, muscovite 15-30%, garnet 2-5%, and minor sillimanite and graphite. The assemblage is thus characteristic of the zone above the breakdown of staurolite, but below the first occurrence of sillimanite plus K-feldspar. The abundant pegmatite segregations that may be a product of partial melting, consist of about equal proportions of quartz and sodic oligoclase with minor muscovite, biotite, and garnet. The schist is commonly rich in biotite at contacts of segregations.

A curious pegmatite dike occurs on the eastern contact of the gneiss of the Tully body. It has the appearance of having been generated by partial melting in the gneiss of the Tully body, and has been injected through the contact amphibolite (note discordant contacts) into the overlying Partridge Formation. Since intrusion the pegmatite and its country rock have been folded in a series of large normal asymmetric folds.

Return to pavement westbound on Route 2 for a short distance.

36.3 Go to far end of grass strip on right beyond one section of guard rail. This is firm in all weather and is the parking spot for STOP 6, which is in a large road cut on Bachelder Road north of Route 2 where there is essentially no traffic. Trail leads from northwest end of grass strip 100 feet to low point of fence. In clear weather WATCH OUT FOR FALLING PARACHUTISTS!

STOP 6. See Figure 17. In this cut is exposed the contact between the Partridge Formation next to the main body of Monson Gneiss and the North Orange band of Monson Gneiss to the east. If the interpretation given under Stop 5 is correct, then all of these rocks lie above the Brennan Hill thrust. The Partridge consists of purplish, rusty-weathering plagioclase-biotite schist with minor garnet and sillimanite and abundant amphibolite layers. The Monson consists of interlayered plagioclase-biotite gneiss and hornblende amphibolite. A few layers contain garnet and anthophyllite that are not characteristic of the Monson elsewhere.

The dominant structural features are gently plunging, open to isoclinal anticlines and synclines parallel to the regional trend of minor folds and lineations believed to have formed during the dome stage. The folds show a normal sense of asymmetry indicating a major synclinal structure to the west, in this case the Williams Pond syncline that runs down the center of the main body of Monson Gneiss. Prominent amphibolite boudins in gneiss and schist have neck lines normal to fold axes suggesting dome-stage north-south extension. In the schist there are several examples of earlier folds with folded axial surfaces.

In the gneiss there are several examples of earlier lineations and folds folded across the later folds. The most spectacular example is at the top of the third gneiss outcrop on the north side of the road. Here a gently plunging late anticline has a biotite lineation parallel to its axis. A prominent earlier lineation is folded across the crest and a related isoclinal fold, plunging southeast, can be seen on the east limb. This earlier fold is clearly a fold in foliation and closely resembles 2nd-stage folds assigned to the backfold stage

in large outcrops in the Quabbin Reservoir area. Even the movement sense of this early fold can be worked out and indicates that structurally higher rocks moved north. This is at least conceptually consistent with the postulated northward flowage of the main body of Monson Gneiss during the backfold stage, which is believed to have involuted the earlier North Orange nappe. This is the best exposure showing fold interference in three dimensions that has been seen in central Massachusetts outside the Quabbin Reservoir area.

END OF TRIP.

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# GEOLOGY OF THE PETERBOROUGH AND CONCORD QUADRANGLES, NEW HAMPSHIRE

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## INTRODUCTION

The Peterborough and Concord quadrangles span a geologic terrane from the eastern portion of the Kearsarge-Central Maine synclinorium into the axial region of the Central New Hampshire anticlinorium where, in the southeastern part of each quadrangle, formations and structures are truncated by the Campbell Hill fault. Southeast of the fault is the Precambrian (?) ( no younger than Early Ordovician) Massabesic Gneiss. Readers of this quidebook are referred to the accompanying article by Eusden on the Gilmanton quadrangle for a concise synthesis of the stratigraphy and structure of central and southern New Hampshire.

A description of the geology of the Peterborough quadrangle has previously been published by Greene (1970), and of the Concord quadrangle by Vernon (1971). Because both studies were undertaken before correlations had been established between the stratigraphy of western Maine and that of central New Hampshire (Hatch, Moench and Lyons, 1983) all of the metasedimentary rocks in both quadrangles were erroneously identified as members of the Littleton Formation. Remapping of the Peterborough quadrangle was undertaken by E. F. Duke (1984) as part of a Ph. D. study. The portion of the Concord quadrangle east of the Weare pluton of Kinsman Quartz Monzonite was remapped by G. I. Duke (1984) as part of an M.Sc. study. The western part of the Concord quadrangle and the eastern part of the adjacent Hillsboro quadrangle were remapped by Carl Hanson and J. B. Lyons (ms.) with the aid of funds from the Office of the N. H. State Geologist.

The mapping in the Concord quadrangle was greatly facilitated through the help of W. H. Vernon, who generously made his manuscript and outcrop location maps available to us. Greene's (1970) geologic map was also very useful, particularly as a guide to the plutons, which allowed a concentration of the mapping effort on the metasedimentary geology.

## STRATIGRAPHIC CORRELATIONS

The Massabesic Gneiss (Figs. 1 and 2) belongs to a widespread terrane characterized by pink orthogneiss and gray migmatitic paragneiss, both with enclaves of calc-silicate , granofelsic, or schistose rocks; all these are cut by a variety of Devonian (?) granitoids. The Berwick Formation, where unmigmatized, is a purplish biotite granofels with discontinuous layers or pods of calc-silicate, and locally, beds of sulfidic schist.

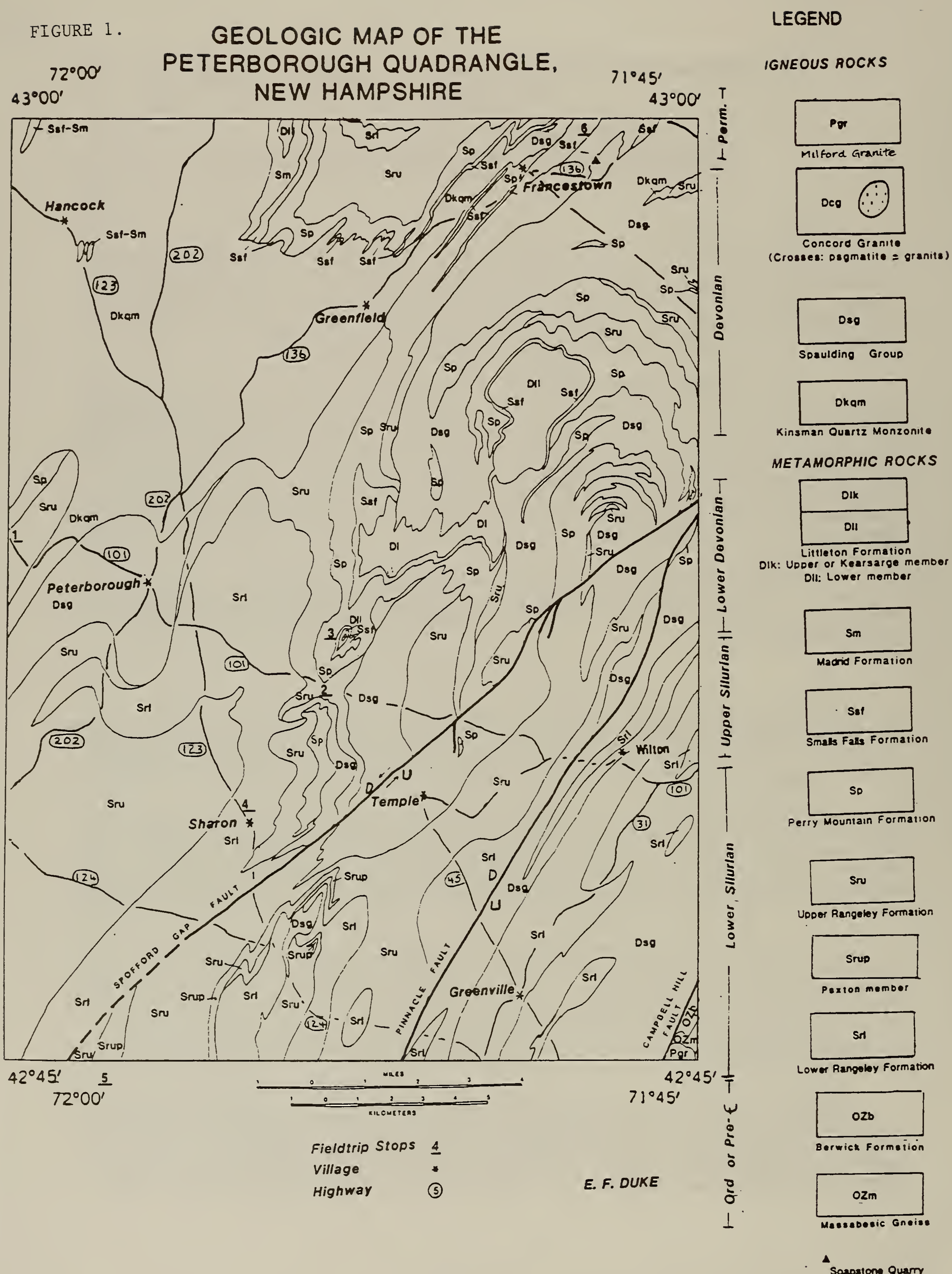
Aside from the plutonic rocks immediately northwest of the Campbell Hill fault, metasedimentary rocks there belong to the Rangeley Formation. The lower Rangeley in this area is a thinly-laminated metapelite, with rare calc-silicate boudins, whereas the upper Rangeley is generally rusty-weathering, with relatively abundant calc-silicate boudins. There are, however, exceptions. In the Gilmanton quadrangle (Eusden, this guidebook) calc-silicate boudins are common in the lower Rangeley, but rare in the upper Rangeley. This is also true in the northeastern part of the Penacook quadrangle (Lyons, this guidebook).

Between the Pinnacle and Campbell Hill faults of the Concord quadrangle, G. I. Duke distinguishes a unit (Grasmere Member of the lower Rangeley) which is more coarsely laminated than the typical Rangeley, carries calc-silicate boudins, and is locally rusty. This corresponds to what, in the Concord quadrangle was mapped as the lowest Littleton (Vernon, 1971) and in the Peterborough quadrangle, the Souhegan Member of the Littleton Formation (Greene, 1970). These rocks are quite similar to Eusden's lower Rangeley. It is conceivable, but certainly not provable, that they could be correlative with the Lower Silurian Greenvale Cove Formation of northwestern Maine, which underlies the Rangeley.

The Perry Mountain Formation is recognizable because it has sharply interbedded metapelitic and quartzite beds (couplets up to 15 or 25 cm. on average), has relatively abundant coticule beds, and is only locally rusty-weathering. Throughout most of central New Hampshire it lacks calc-silicate boudins, but in eastern New Hampshire (Eusden, this volume) boudins are present. Parts of what was mapped by Greene (1970) as the Crotched Mountain Member of the Littleton Formation is partly Perry Mountain and partly Rangeley Formation. Nielson's (1981)

FIGURE 1.

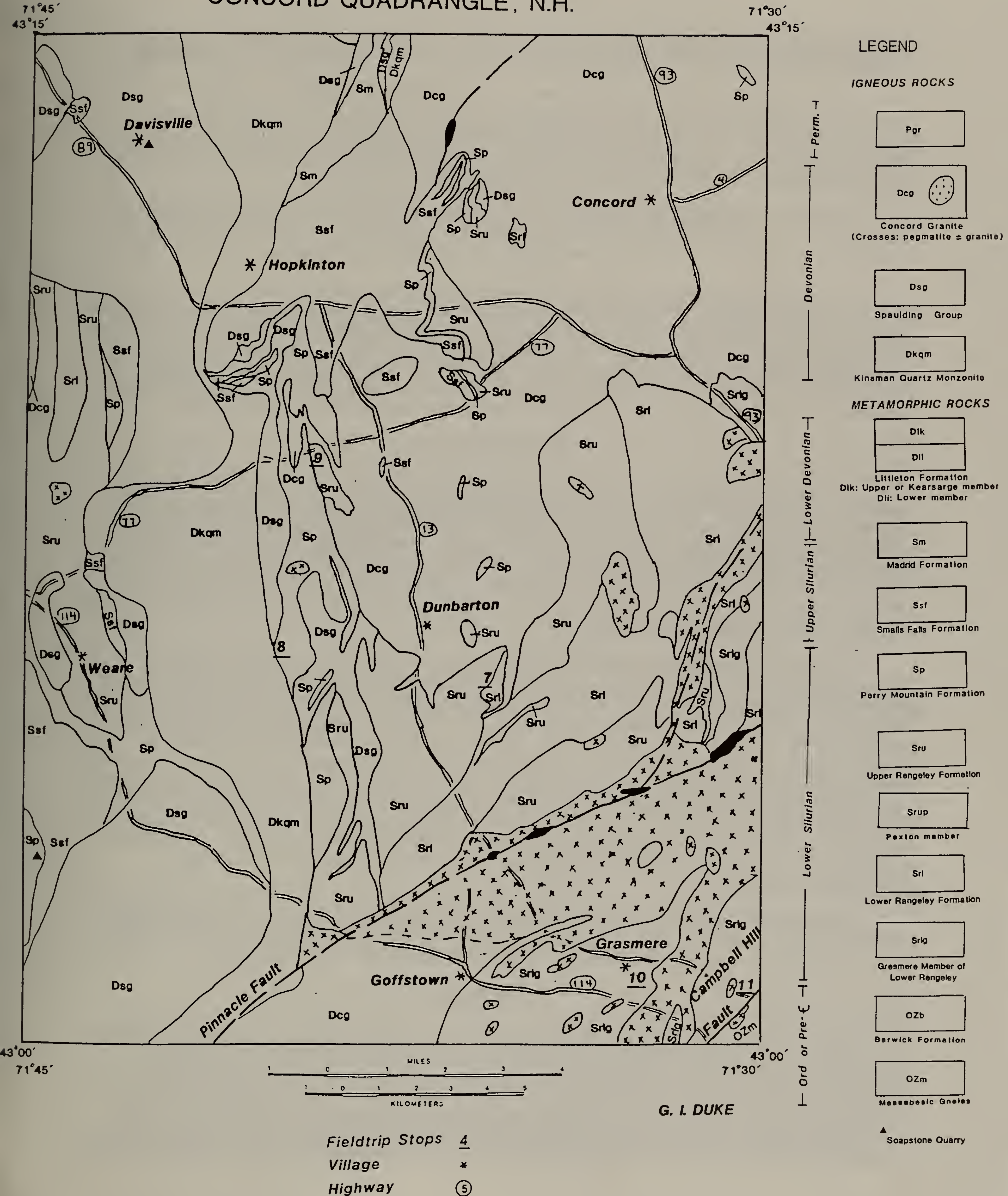
# GEOLOGIC MAP OF THE PETERBOROUGH QUADRANGLE, NEW HAMPSHIRE




Fieldtrip Stops 4
Village *
Highway (5)

E. F. DUKE

FIGURE 2. GEOLOGIC MAP OF THE CONCORD QUADRANGLE, N.H.



Crotched Mountain Formation is the Perry Mountain, but his rusty Crotched Mountain Member is the upper Rangeley. Englund's (1976) Roundtop Quartzite member of the Littleton is a sub-unit of the Perry Mountain.

The most useful and distinctive stratigraphic marker in central New Hampshire is a strongly rusty-weathering interbedded pyrrhotitic calc-silicate and white quartz-muscovite-graphite schist. This was initially identified by Greene (1970) as the Francestown Member of the Littleton Formation. Englund (1976) mapped a similar formation in the Holderness quadrangle as the Clay Brook Member of the Littleton. It is now clear from regional mapping that both these units are the Upper Silurian Smalls Falls Formation.

A purplish biotite granofels with distinctive calc-silicate boudins and, locally, some rusty-weathering metapelites was first mapped by Lyons (ms.) in the Mt. Kearsarge quadrangle as the Warner Formation, but it is now recognized as identical stratigraphically and lithologically with the Madrid Formation of northwestern Maine.

The Littleton Formation in central New Hampshire consists of gray schists. Generally, but not always, the lower Littleton is nondescript, whereas the upper Littleton is characterized by excellent graded bedding. There is no Littleton in the Concord quadrangle. In the Peterborough quadrangle, graded bedding is characteristic of the Littleton. Calc-silicate boudins are very rare, but a few have been observed, which may lead to some ambiguity in distinguishing boudin-bearing lower Rangeley from the Littleton. The local stratigraphic sequence, and the characteristic "slow grades" of the Littleton will help to resolve the uncertainties. Perry Mountain rhythmites are "fast graded", and often difficult to read for facing directions. This is generally not true of the Littleton.

## PLUTONIC ROCKS

The oldest and most distinctive of the New Hampshire Series plutonic rocks is the Kinsman Quartz Monzonite, characterized by its very large megacrysts of potash feldspar and, locally, by abundant garnet. The Peterborough quadrangle contains a southeasterly segment of the very large Cardigan pluton, which extends southerly from here for another 40 miles as the Coys Hill Granite of Massachusetts. In the western Concord quadrangle the Weare pluton of Kinsman Quartz Monzonite forms a separate body which, however, comes within a few miles of joining the Cardigan pluton both in the Mount Kearsarge and in the Monadnock quadrangles. It is likely that the two plutons were at one time parts of the same sheet ( Nielson and others, 1976) and that their current separation is due to folding and subsequent erosion.

Younger than the Kinsman and older than the post-tectonic Concord Granite is a group of plutons belonging to the Spaulding Intrusive Suite (Dsg on the maps). These are late-tectonic, weakly to strongly foliated, and petrographically diverse. Most of the rocks are biotite tonalites or granodiorites (some with garnet), but they range into two-mica granites. The Hopkington pluton (Dsg in the northwestern part of the Concord quadrangle) is more mafic than other Spaulding intrusives, and is a hypersthene-biotite quartz diorite with some phases of hornblende gabbro. Distinctions between the peraluminous granitoids of the Spaulding and Concord Intrusive suites are difficult to make, and are based largely on field and microscopic evidence as to whether the plutons show tectonic (Spaulding) or flow (Concord) foliations. As is to be expected where there is no closely controlled isotopic dating, interpretations may differ. This is true, for example, of plutons along the boundaries of the Concord (Vernon, 1971) and Hillsboro (Nielson, 1981) quadrangles. Similarly Greene (1970) and E. F. Duke (1984) do not classify the granitoids in the same way within the Peterborough quadrangle. Until more isotopic dating is available, uncertainties will remain.

Despite the fact that it is the type pluton for the Concord Plutonic Suite, efforts to determine a reliable isotopic age on the Concord Granite stock have been hampered by inheritance problems in the zircons. A probable age (J. N. Aleinikoff, verbal communication, 1986) is approximately 365 Ma. (Later Devonian).

## SOAPSTONES

In the nineteenth century soapstone was quarried at four locations in central and southern New Hampshire. One of these is in the northeastern Peterborough quadrangle (Francestown), two in the Concord quadrangle (Davisville, and Mt. Misery in Weare) and one in the Penacook quadrangle (Canterbury). Their geochemistry (D. R. Nielson, 1974) differs from that of alpine ultramafics, and their geologic setting is puzzling. One of them (Davisville) is a very small xenolith in the Spaulding, another (the largest; Francestown) is at a contact between Spaulding and the Perry Mountain, a third (Mt. Misery in Weare ) has Perry Mountain as one wallrock and quartz monzodiorite at the other; the latter could be an apophysis from a Spaulding pluton nearby, but the fourth and second

largest (Canterbury) is along the contact between Madrid and Smalls Falls. Bulk geochemistry (D. R.Nielson, 1974) implies that the protoliths of the soapstones were mafic gabbros, possibly with ultramafic layers. Mineralogically they consist chiefly of talc and chlorite, with varying amounts of hornblende or pargasite, actinolite, anthophyllite, biotite, olivine (rare), pyrrrhotite and magnetite.

The mechanism of emplacement is occult. The Canterbury occurrence as well as density considerations make it unlikely that the mafic rocks were floated into position during the emplacement of the Spaulding intrusives. A more plausible scenario is that there may be a series of pre- or syn-metamorphic and as-yet-unmapped duplex-type faults, some of which step downward into sheets of mafic igneous rocks. Blocks of these mafic rocks would have been dragged into their present positions during faulting, before the Spaulding plutons exploited the same faults as passageways for emplacement, and also before the completion of regional metamorphism.

## FAULTS

Another clue to the hypothesis that there may be unmappped pre- or syn-metamorphic faults in New Hampshire is suggested by the results of the COCORP traverse in New Hampshire (Ando and others, 1984). The line crossed the region of the Concord and Penacook quadrangles close to their mutual contact, and shows a number of strongly defined reflectors at depth, stepping downward easterly in a pattern analogous to those of fault duplexes. Whether the reflectors are all faults, or all intrusive sheets,or some combination thereof ( regarded by us as most likely) cannot be proven at present.

Several aspects of the mapped geology are also difficult to explain unless these are pre- or syn-metamorphic faults sub-parallel to stratigraphic boundaries. In the Peterborough quadrangle, for example, the Madrid formation is present in the northwestern part of the quadrangle, but throughout the central area of the quadrangle the Littleton Formation is in contact with a thin strip of the Francestown. In the Hillsboro quadrangle to the north (Nielson, 1981) a belt of the Madrid thins southerly from the Mt. Kearsarge quadrangle over a distance of four miles, is missing for the next five, and is present for the next 10 as it passes into the northwestern Peterborough quadrangle A Littleton-Smalls Falls contact is present in the Gilmanton quadrangle (see Eusden, this guidebook) , but there the lower contact of the Littleton contains clasts of the Smalls Falls ("Wild Goose Grits"), and may be an unconformity.

An analogous problem exists in the Concord quadrangle, but involves different stratigraphic units, so the difficulty is not confined to the contacts of the Littleton. North of Weare, the Smalls Fall and upper Rangeley Formations are juxtaposed for a few miles, with the Perry Mountain missing. North of this, across a break interposed by intrusives, the same units are in contact for another mile, but a wedge of Perry Mountain is present for the next two miles, and then is missing for another mile before this contact is eliminated by an intrusive. We do not yet have sufficient control on the premetamorphic faults to show their distributions accurately on maps, nor to draw them in cross sections.

Faults of a different nature are displayed in a series of silicified zones traversing the Peterborough and Concord quadrangles. These are, from west to east , the Spofford Gap, Pinnacle, and Campbell Hill faults, and an unnamed fault in the northern Concord and southern Penacook quadrangles. The Campbell Hill fault joins the Norumbega fault toward the northeast in Maine.

E. F. Duke (1984, p. 72) has shown that the Spofford Gap fault, which is a splay off the Pinnacle fault, has a probable left-lateral displacement of four miles (6 km), based on palinspastic reconstruction of offset contacts, and the distribution of fault splays and kinks. Offset on the Pinnacle fault is less well constrained, but may be left-lateral, with the west-side-down motion. Juxtaposition of Silurian-Devonian rocks on the west against Precambrian rocks on the east also argues west-side-down motion along the Campbell Hill fault.

The timing of these post-tectonic faults is somewhat uncertain, except that it must be younger than the Acadian orogeny, because Acadian structures are truncated by the faults, and older than a syenite dike which cuts the Campbell Hill fault in the northeastern Milford quadrangle, and has a fission-track zircon age of160 Ma (Aleinikoff, 1978, pp. 116-117). The timing can probably be more closely restricted, however, because there are numerous pegmatites and granites in or close to the trace of the Campbell Hill and Pinnacle faults which are probable "stitching" plutons. One of these, the Barrington pluton or Center Strafford pluton of the Gilmanton and Wolfeboro quadrangles has been interpreted as a stitching pluton, has S-C fabrics, and also a monazite Pb/Uage of 364 Ma. (Eusden, 1988). Very likely, initial motion on the Campbell Hill fault was slightly after the termination of the Acadian orogeny. Motion on the fault was probably active over a long period of time, however, because the

Permian (275 Ma; Aleinikoff, 1978) granite at Milford in southern New Hampshire is another such stitching pluton, strung out along the trace of the Campbell Hill fault. As pointed out by Don Wise (verbal communication, 1988) some of silicified zones along these faults have open-space vugs and evidence of repeated opening and infilling. These features are very unlikely unless the rocks are in the brittle zone ( i.e. at a shallow level), so some of the silicification and faulting may be as young as Early Mesozoic age,but prior to 160 Ma age of the syenite dike which cuts the Campbell Hill fault.

## METAMORPHISM

Within the Peterborough quadrangle, all the metasediments are in the sillimanite-muscovite subfacies of the amphibolite facies, and lie east of the $Al_2SiO_5$ isobar. The rocks are retrograded locally and the single observed sample containing staurolite (E. F. Duke, 1984) is from a retrograded metapelite. Calc-silicate rocks contain varying assemblages of the minerals grosssularite- diopside-actinolite-zoisite-biotite or phlogopite- K feldspar- plagioclase-quartz - carbonate.

The Concord quadrangle, by contrast, shows a good example of a regional "hot spot" both mineralogically, and in the biotite-garnet core and rim isotherm pattern (Fig. 3; G. I. Duke, 1984). Pressures as calculated by the garnet-biotite-$Al_2SiO_5$ -quartz geobarometer range from 3.4 to 4.2 kb.in 6 samples (Duke, op. cit). Their average (3.8 kb) is close to that to be anticipated for an area lying east of the $Al_2SiO_5$ isobar . The cause of the hot spot cannot be directly related to any known feature of the mapped geology.

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## ITINERARY

Field trip commences in Keene, and prodeeds easterly for 12 miles on Rte 101 through Marlborough and Dublin to the intersection of Rte 137 with Rte 101. This is mile 0.

Road Log (in Miles)

0.0 Junction of Rte. 137 and 101. Proceed easterly on 101.

1.6 STOP 1 Excellent exposures of the rusty-weathering upper unit of the Rangeley at the boundary between the Monadnock and Peterborough quadrangles. The unit dips steeply west-northwest and consists of a variety of massive, laminated, bedded, or banded schists and granofelses with abundant calc-silicate beds and pods.

4.5 Junction with Rte. 202. Stay on 101.

8.6 CAUTION. Left turn into Miller State Park at the top of a long hill. Park cars, and walk to highway.

STOP 2. Rocks along the highway are interpreted as representing the uppermost portion of the Rangeley Formation at or near its transition into the overlying Perry Mountain Formation. These strata include a heterogeneous succession of northwesterly-dipping , variably rusty-weathering rocks with locally abundant calc-silicate pods or boudins Dikes and sills of pegmatite-aplite and biotite-muscovite granitoids are common, probably related to the Devonian Spaulding (?) Whitcomb Peak granitic pluton one-half mile to the east. Isoclinal F1(?) of F2(?) folds with west-over-east movement are common here, as are later open F4(?) folds with associated crenulation cleavage. F4 axes plunge north-northwest; F1-F2 axes northwest or southwest.

Not to be seen on this excursion, but of interest from the perspective of the regional geology are some analyzed rhyodacitic metatuffs in the upper Rangeley of the northeastern Peterborough quadrangle ( E. F.Duke, 1984, p. 18-19 ). Moench et al ( 1987) report similar extensive felsic tuffs in the Rangeley and Perry Mountain Formations of the Piermont allochthon in western New Hampshire.

Walk up the hill to outcrops in the clifs on the west side of the road, which are in the Perry Mountain formation. There are about 120 m. of resistant interstratified quartzite and pelite in "fast-graded" beds 1/2 to 3 inches ( 1 to 8 cm.) in average thickness, weathering to gray nonrusty outcrops, and facing upward. These are best seen about 1/2 mile from the park entrance at the first switchback, but parking there is difficult. Calc-silicate pods are not conspicuous but occur sporadically throughout this portion of the Perry Mountain. The bedding here dips gently northwesterly, but with many undulations.

Return to cars and drive to the top of Pack Monadnock Mountain.

10.1 Summit parking area. Space is limited

STOP 3 About two-thirds of the way to the summit the Perry Mountain gives way to rusty outcrops of the Smalls Falls. Unlike the Smalls Falls to the west and north these rocks are not calcareous, but do resemble much of the Smalls Falls of the type locality in Maine, as well as similar rocks in Massachusetts which have been mapped as the Smalls Falls. Though somewhat similar to Rangeley lithologies at the base of the mountain, these rocks lack the calc-silicate pods which are typical there, and also on the basis of graded bedding in the Perry Mountain lie stratigraphically above, rather than below that formation. The Smalls Falls is interpreted to have undergone a facies change from metapelites on the southeast to calcareous lithologies on the northwest.

Immediately north of the parking lot and west of the small shelter is what appears to be inverted, thinly-bedded Perry Mountain facing downward toward the rusty Smalls Falls.

Walk north-northwest from the parking lot along the Wapack Trail.The path follows rusty schists for several hundred feet and then crosses into massive non-rusty outcrops across a sharp contact. Along western slopes of this open ridge are good exposures of garnetiferous "coticule" lenses, interpreted to be unis of a highly attenuated infold of Lower Devonian Littleton Formation as exposed on North Pack Monadnock Mountain. The complex outcrop pattern here is believed to be result primarily from interference of north-northwest trending F1 folds by north-northeast-trending F4 folds.

Retrace route to entrance of Miller State Park

11.6 Right turn onto Rte. 101 toward Peterborough

14.3 Left turn on Rte. 123 toward Sharon. DANGEROUS TURN

17.6 Park cars off the highway. Shoulder is very narrow here, so be careful

STOP 4. Several outcrops in the woods east of the road display features typical of the lower Rangeley. The rocks have a laminated or thinly-bedded aspect caused by compositional layering. Although only a few small calc-silicate beds or lenses occur at this location, elsewhere conspicuous calc-silicate pods may measure several meters in plan view. The mineralogy here consists of quartz, plagioclase, biotite and muscovite with subordinate sillimanite, and locally abundant quartz-feldspar streaks and lenses. Intruding the metasediments are sheets of foliated Spaulding (?) intrusives.

The massive to thinly-laminated character of the lower Rangeley distinguishes it from either the more uniformly and somewhat thicker-bedded Perry Mountain, or the even more thickly and variably graded-bedded Littleton. A minimum estimate for the exposed thickness of the lower Rangeley in the Peterborough quadrangle is 400 meters; however, the lower contact has not been exposed.

Proceed southerly on Rte. 123.

17.8 Sharon

19.2 Right on second crossroad.

20.2 Right on Rte. 124.

23.3 Left on Prescott Road, just beyond Millipore Filter factory on left..

25.6 CAUTION. Left at Cathedral of the Pines sign.

27.0 Bear right at T.

27.5 Left on Old Ipswich Road.

28.9 Left on Perry Road.

30. Right on unmarked road. This is North Road, a half-mile north of the hamlet of East Rindge in the Ashburnham (1:24,000) quadrangle.

30.5 Park at large gravel pit on the right.

STOP 5. Small road cut near the base of a utility pole. Despite its inconspicuous nature, this is a significant outcrop for tying some of the Massachusettts formations into those of New Hampshire. The distinctive banded calc-silicate granofels provides the best outcrop of Paxton-type lithologies in the area. Stratigraphically the Paxton apparently lies just above the transition from the lower (non-rusty) into the upper (rusty) member of the Rangeley. Two miles to the east, Peterson (1984) maps a similar Paxton granofels into the Srup unit of Duke's (1984) Peterborough quadrangle. This would establish, as well as any other available evidence, an Early Silurian age for the Paxton.

This member of the Paxton is characterized by non-rusty light-hued outcrops composed of contrasting 1-5 cm. bands of purplish quartz-plagioclase-biotite granofels and light green or pink calc-silicate granofels. The calcareous belt is only 6 meters thick where it is exposed in the Peterborough quadrangle. Despite this apparent tie-in of the Rangeley and Paxton Formations it is only fair to note that Duke's (1984) lower Rangeley east of here is mapped as Littleton by Peterson (1984) in the Ashburnham-Ashby quadrangles.

Turn around and retrace route through Sharon toward Peterborough.

46.7 Dangerous Intersection. Cross Rte 101, continuing on Rte 123

49.2 Junction of Rtes 123, 202, and 136. Sharp right turn on Rte 136

55.2 Greenfield

59.7 Francestown . Right turn on Rte 136

61.3 Left turn on Bible Hill Road

62.3 STOP 6 Type outcrops of the Francestown Member of the Littleton Formation of Greene (1970); ( = Smalls Falls Formation). The Smalls Falls Formation in the Peterborough quadrangle is a thin unit that crops out in several structural belts. In the northern portion of the quadrangle it consists chiefly of sulfidic calc-silicate granofels containing varying proportions of quartz, phlogopite, microcline, calcic plagioclase, actinolite, diopside, zoisite, sphene and pyrrhotite. These are interstratified with white graphite-quartz-microcline-muscovite-phlogopite schist. Some of the lithologic members in the Fitch Formation of the Lovewell Mountain quadrangle to the northwest greatly resemble the Smalls Falls, and a correlation is likely. Bedding parallels foliation here and averages N.31 E, 26 NW. The Smalls Falls Formation in the Peterborough quadrangle is very thin, perhaps only 10-50 m. thick in some places. The outcrop width here is misleading, because at these outcrops the Smalls Falls is thought to lie in a structural keel.

Turn around and head toward Rte 136

63.1 STOP 6A (?) Francestown Soapstone Quarry As described by D. R. Nielson ( 1974) the Francestown soapstone is primarily a tremolite-talc- phlogopite-chlorite rock and thus, unlike the other soapstone bodies of central New Hampshire the protolith may not originally have been a mafic igneous rock but rather, as proposed by Greene (1970, p.58), an argillaceous dolomite. The alternative of a metasomatized mafic rock cannot be ruled out, because portions of the soapstone contain remnant hornblende or actinolite with a lamellar texture, which suggests the uralitization of original augite. The northeast wall of the quarry is the northwesterly-dipping Perry Mountain Formation, and the southeast wall is a biotite-rich blackwall. Further southeast are Spaulding granitoids.

63.5 Left onto Rte. 136.

69.3 Junction with Rte. 77. Continue southerly on Rtes. 136 and 77.

69.6 Left onto Rte 13 in New Boston

76.0 Right on Rtes. 13 and 114 toward Goffstown

76.3 Left on Rte. 13 toward Dunbarton

81.8 Left toward Clough State Park

83.1 Left toward Everett dam

84.7 Everett Dam. Park cars

STOP 7 Perry Mountain Formation, well exposed on the northeast bank of the dam, and particularly in the spillways, showing typical "fast grades" , but also, in the spillways, some unusually rusty outcrops of uncertain affinity (Rangeley(?), Perry Mouintain(?), or Smalls Falls (?)). The structure and lithology in the spillway outcrops is complex, with recumbent folds both in the Perry Mountain and in some of the sheets of Spaulding (?) granitoids intruding it. In the granites a sub-horizontal axial plane cleavage has been developed, indicating that they are partly syntectonic. Interestingly, fold axes directions in the recumbent folds vary by as much as 90 degrees. The Kuncanowet Hills to the east have as many as 10 repetitions of the Rangeley-Perry Mountain contact.

Raymond Cliffs to the west across the dam is the well-exposed eastern basal contact between the Kinsman Quartz Monzonite of the Weare pluton and the Perry Mountain. Here you look under this sheet-like pluton at its semi-concordant floor of Perry Mountain.

Retrace trip toward Rte. 13.

86,7 Right at T.

88.0 Left on Rte. 13.

88.3 Right at Dunbarton common on Robert Rogers Road.

89.2 Right on dirt road (a.k.a. Leg Ache Hill Road).

90.1 Find a spot to park your car off the road, and walk down the hill to the powerline.

STOP 8 Lower and upper Rangeley outcrops. The lower Rangeley outcrops south of the road are characteristically thinly laminated (0.1 to 2 cm. average) gray metapelite with sillimanite, biotite, garnet, plagioclase and quartz. With the exception of the Grasmere member (q.v.) there are no calc-silicates in the lower Rangeley of the Concord quadrangle. As is characteristic of much of the lower Rangeley in central New Hampshire, migmatization appears to be incipient, with the development of anatectic (?) quartz-feldspar pods and veinlets. One sample from this series of outcrops has the assemblage qtz-sill-bio-gar-cord-plag-K-feldspar, which places it in the granulite facies. It lies in the center of a "hot spot" (Fig. 3) with a gar-bio core temperature of 714°-729° C, and a pressure of 3.8 kb (G. I. Duke, 1984).

The upper Rangeley outcrops north of the road are the characteristic rusty-weathering qtz-bio-plag schist with calc-silicate boudins. Muscovite in some of these rocks is likely to be retrogressive.

Return toward Dunbarton.

91.0 Left turn.

92.3 Right on Rte 13.

95.5 Page's Corner. Left ( westerly) on Rte. 77.

96.6 Park cars.

STOP 9. Upper Rangeley formation. Lithic types characteristic of the upper Rangeley occur here.The rock is characteristically a qtz-bio-plag-sill schist with lesser amounts of gar-musc-K fs and pyrrhotite. Upper Rangeley is estimated to be a minimum of 700 km. thick in the Concord quadrangle. Sheets of foliated Spaulding (?) granodiorite cut the outcrop, as does a camptonite dike (50% carbonate replacing augite, 40% plagioclase, kaersutite, and accessories).

Reverse direction on Rte. 77.

97.7 Right on Rte. 13.

101.3 Dunbarton

105.5 Left on Page Hill Road

107.9 Left at fork. Locust Hill Road

108.7 Left at fork, and through the hamlet of Grasmere on Goffstown Back Road

109.2 Left at fork.

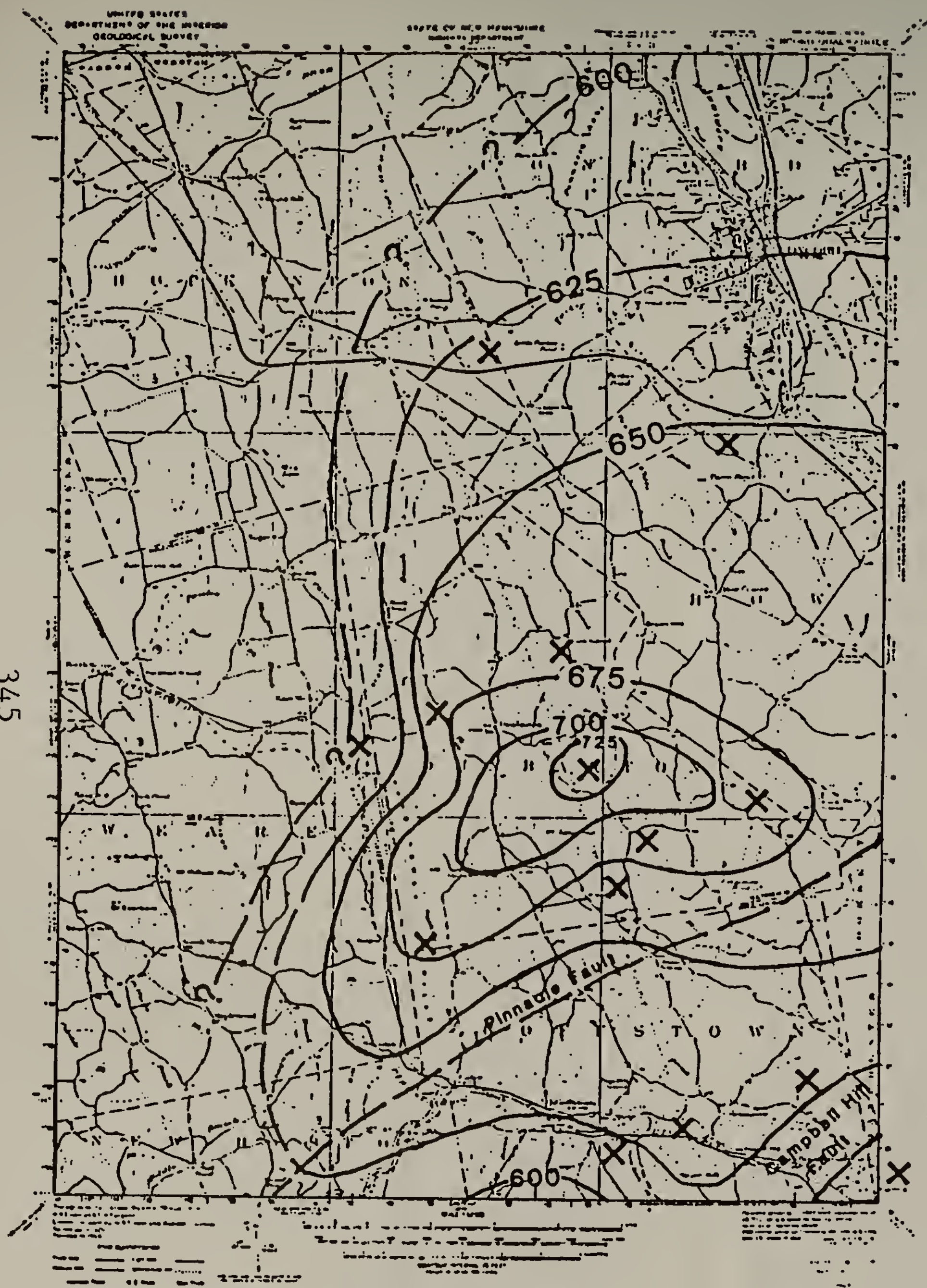
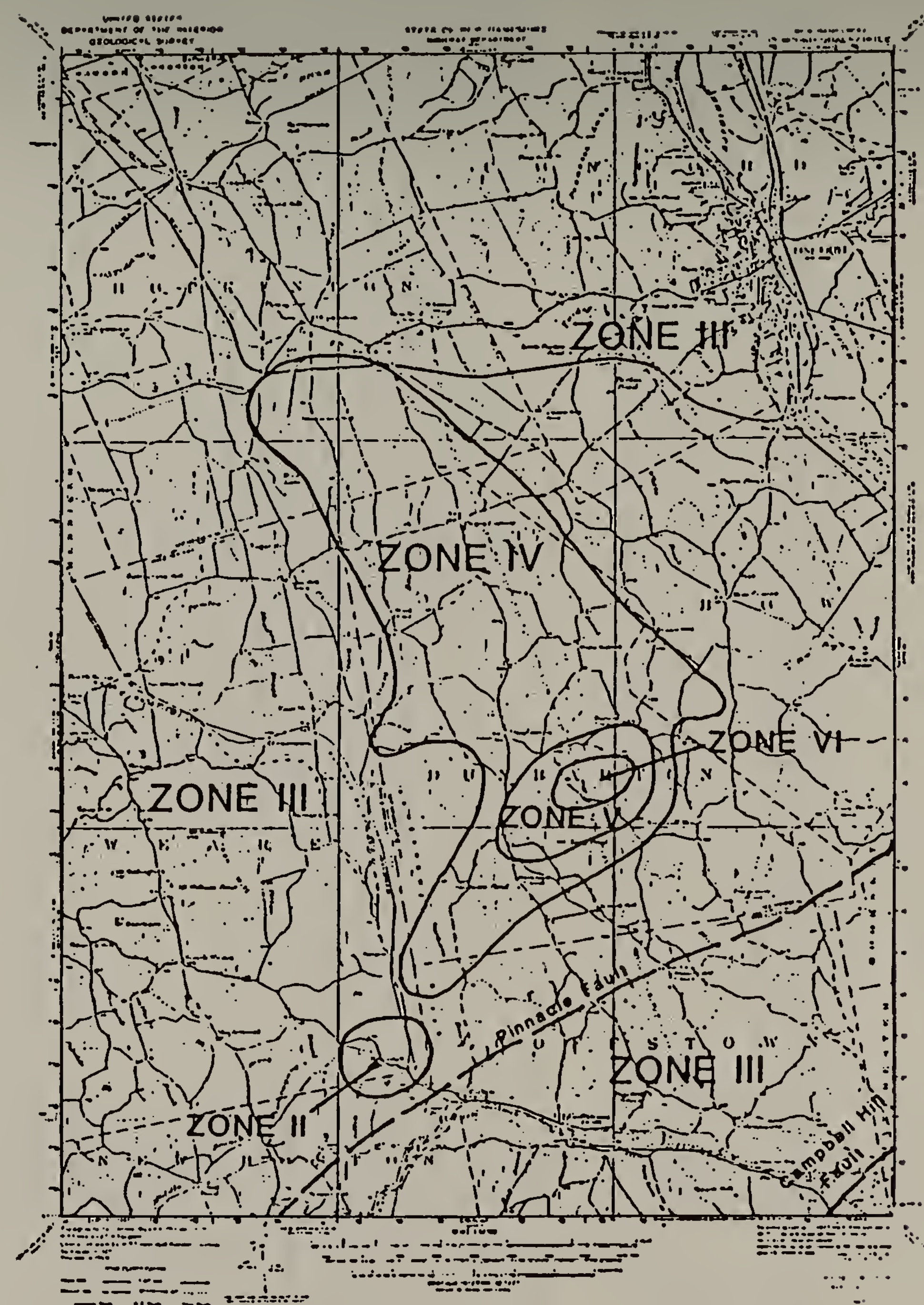



Figure 3: Isotherm map of garnet core temperatures (°C) eastern Concord Quadrangle




Isograd map of the eastern Concord Quadrangle

ZONE II

STAU-SILL-MUSC

ZONE III

SILL-MUSC

ZONE IV

SILL-MUSC K SPAR

ZONE V

SILL-K SPAR

ZONE VI

GAR-SILL-CORD-K SF

109.6 Right at Diamond Road, and park cars at powerline. Note thevery rusty outcrop on Goffstown Back Road, just prior to making this turn.

STOP 10 More thickly laminated metapelite than is characteristic of the lower Rangeley. This feature, the interbedding with very rusty metapelite, and the occurrence of calc-silicate boudins in these rocks led G. I. Duke to separate them from the lower Rangeley as the "Grasmere member". Prior to this, Greene (1970) had labelled them "Souhegan Member of the Littleton Formation", but our current consensus if that they are a facies of the lower Rangeley. Outcrops here are warped by F4 (?) folds bearing 210, and plunging 20 SW.

Continue along Diamond Road.

109.9 Right at T

110.2 Tightly folded sill-bio-gar-plag-musc-qtz-plag "Grasmere",with calc-silicate boudins. Note the development of leucosomes, which is characteristic of the lower Rangeley. Return to Goffstown Back Road.

110.9 Right turn toward Manchester

112.9 Pegmatite outcrop ( typical when nearing the Campbell Hill fault) recently blasted away

113.1 Park cars in field near radio tower.

STOP 11. Campbell Hill fault outcrops north across the road show quartz veining and silicification of pegmatite. In the woods toward the west are screens of the Rangeley (or Grasmere) within the pegmatite, just as there are in the area undergoing development north of the recently blasted pegmatite outcrop.

Massabesic Gneiss crops out along low ledges slightly north of the radio tower, but far better outcrops of the Massabesic (both paragneiss and orthogneiss) have been uncovered in a field just east of Greatstone condominiums, where there is a recently cleaned-off glacially or stream-scoured outcrop. Slightly further east , near the entrance to the Holy Trinity cemetary north of the road are outcrops of well-layered biotite granofels, calc-silicate granofels and interlayered rusty schists of the Berwick Formation .

END OF FIELD TRIP

Followed toward the east, Goffstown Back Road intersects I-293 west of the Merrimack River in Manchester. I-293 joins I-93 both to the north and to the south. For another look at the Campbell Hill fault, the trip can be continued as follows, by proceeding easterly along Goffstown Back Road.

114.6 Right on Montgomery Street

115.0 Right on Kelley Street

116.2 Right on Rte. 114 A.

118.7 Left on Rte. 114.

119.1 Park car . Cross to rusty outcrop on south side of road.

STOP 11A. Pegmatite and silicified zone, similar to those west of Greatstone condominiums. These are cut by a pyritic rusty-weathering syenite dike on whicn Aleinikoff ( 1978) determined a 160 Ma fission-track age on zircon. This would be a minimum age for the last motion on the Campbell Hill fault.

Continuing southeasterly on Rte 114 will lead to an intersection with I-293 in Bedford, west of Manchester, and at the Merrimack River.

# RELATIONSHIP BETWEEN BEDROCK GEOLOGY AND GEOMORPHOLOGY IN PORTIONS OF THE COLD RIVER AND ASHUELOT RIVER WATERSHEDS, SOUTH-CENTRAL NEW HAMPSHIRE.

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In north-central Maine, Hanson and Caldwell (1988) have found a clear relationship between bedrock geology and topgraphy. In particlular, in the Greenville-Rockwood area metasandstones and felsites form long, continuous ridges. The highest terraine is composed of hornfels in contact aureoles near Devonian plutons. The erosion of granularly disintegrated Acadian plutonic rocks has formed numerous lake basins and low hills. Slatey rocks have also been weathered and eroded to low topgraphy. In southwestern New Hampshire, with higher grades of regional metamorphism, the relationship between bedrock geology and topography is more complex than in Maine. The Acadian plutonic terraines, however, are similar in the two areas.

The field area extends through portions of three major structural belts, the Connecticut Valley-Gaspe Synclinorium (CVGS) on the west, the Bronson Hill Anticlinorium (BHA), and the Kearsarge-Central Maine Synclinorium (KCMS). These fold belts were formed during the Acadian Orogeny. Pre-Silurian rocks in the BHA generally contain evidence of multiple episodes of deformation, while Silurian and Devonian metasediments are generally less complexly deformed. The CVGS in this area is underlain by the Devonian Littleton Formation. The Littleton has a variety of topographic expressions, underlying the Connecticut Valley (Stop 1), is draped over the gneiss domes in the BHA (Stops 2 and 5), and holds up the summit of Mt. Monadnock in the KCMS. This variation in resistance to erosion is related to degree of regional metamorphism and to the thermal effects of Devonian plutons. The BHA is characterized by nappe structures which are cored by Oliverian (Ordovician) gneiss and covered by a variety of Ordovician through Devonian sedimentary and volcanic rocks (See Chamberlain, Thompson and Allen's (A-2) guide in this volume for a new interpretation of the nappe stratigraphy in this area). The Alstead dome (Stop 5) has been breached, with overlying units dipping symmetrically away from the gneiss core. The Ammonoosuc volcanics have locally been weathered to orange and red, clay-rich saprolite, some exposures of which are under till (Stop 4). This saprolite is considered to be Tertiary in age. The Beryl Mountain pegmatite forms a prominant knob within the valley of the Cold River (Stop 6). This and other pegmatites in the area are likely late Paleozoic in age, as other New England pegmatites are now believed to be. East of the gneiss domes there are large areas underlain by the Devonian Kinsman porphyritic granite. The Kinsman in this area underlies long narrow valleys, such as the headwaters of the Ashuelot and Contoocook Rivers (Stops 7 and 8). The highest elevations in the field area are underlain by rocks assigned to the Silurian Rangeley Formation (Stop 8).

Landslides and catastrophic floods continue to modify the valley of the Cold River. Although houses were built by the early settlers on the floodplain, they were reached only by roads descending from the uplands. Some roads that run along the valley bottom were constructed as recently as 1914.

## REFERENCE

Hanson, Lindley S. and Caldwell, D.W., 1988, Relationship between bedrock geology and geomorphology in the mountainous areas of northcentral Maine, *in* Tucker, R. and Marvinney, R., *eds.*, The Geology of Maine: Maine Geological Survey, Augusta, Maine, .

## ITINERARY

The trip begins in the Keene State College Commons parking lot. No mileage is given until North Walpole, New Hampshire. Exit parking lot and turn right onto Main Street. At first traffic light, turn right onto Route 12. At third traffic light, Route 12 turns right again. Follow Route 12 for about 20 miles to North Walpole.

Miles

0.0 The mileage of the trip will begin near the mouth of Cold River, at the junction of Routes 123 and 12 Continue north on Route 12, with the Connecticut River on the left and Fall Mountain on the right.

0.1 Mouth of Saxton's River and USGS Connecticut River Gauge on left.

0.8 Turn left at traffic light. Roundhouse of Green Mountain Railway on right.

1.0 Oldest canal in the United States.

1.1 Bellows Falls, Vermont. Turn left onto Westminster Street.

1.4 Junction with Route 5 and 121 at traffic light.

Continue ahead on Route 121.

2.3 Turn left onto unnamed street.

2.4 Straight ahead on dirt road.

2.7 STOP 1. Pull over and park on left side of road. Descend bank of Saxton's River on right. Outcrops of Devonian Littleton Formation are regionally metamorphosed to staurolite grade. First turbidite beds encountered dip to the south and are upside down. On the cliff immediately upstream, similar units also dip to the south and are right side up. The stream course here is roughly controlled by the strike of the beds. Sandy portions of turbidite beds have well-developed climbing ripples. Some sandy layers in the turbidites are disturbed and broken, indicating soft-sediment deformation.

Return to cars and continue along dirt road.

3.2 Turn right on Route 121.

3.7 At light, straight ahead on Westminster Street.

4.0 Turn right at corner.

4.3 Turn right onto Route 12 at traffic light.

5.2 Turn left onto Route 123.

5.4 Cross Cold River.

5.7 Turn left again on Route 123 and begin climb up foreset slope of Lake Hitchock delta.

6.6 STOP 2. Lake Hitchcock delta surface. Foreset beds are usually visible in pits below. Most of the coarser topset beds have been mined away and operator has had to crush stone to make up proper aggregate sorting. Stone used in crushing operation is Devonian Bethlehem gneiss quarried a short distance up the road. Fall Mountain nappe across the valley.

Return to cars and continue up Route 123.

7.9 Cross Cold River and turn right on Route 123 at blinking light.

9.7 Turn right at next blinking light on Route 12 and123. Cross Cold River.

10.3 Intersection of Route 123A. Bear right on Route 123 and 12.

10.9 Pull into rest area on right and make U-turn, going back north on Route 123 and 12.

11.3 STOP 3. Pull over on right shoulder and park. Bellow's Slide on Great Brook. See Ridge (B-4, this volume) for a detailed description of this section. Lake sediments are underlain by tills with inderbedded silt. Possible lower till here. This is the largest of a number of slides and flows that continue to modify the valley walls of the Cold River and its tributaries, particularly in the middle third of the watershed. Other slides have and are diverting the Cold River from its course. Commonly slides bring trees into the river which form debris dams, sometimes augmented by beaver works and always by ice jams, which causes the river to move its channel.

Return to cars and continue north.

11.5 Turn right on Route 123A.

13.8 Cross Cold River. There are numerous active alluvial fans in this area that have built onto the flood plain.

14.4 Bear left.

14.6 STOP 4. Park on right of road. Walk up small gully on left. Blue tubing is used to gather maple sap from up the hill. Saprolite developed on Ammonoosuc volcanics. Other saprolite exposures occur up this gully and along other tributaries in this section of the Cold River. Original section exposed in catastrophic flood in 1986 showed till overlying weathered diamict, then saprolite, with bedrock at the base. Exposures of Ammonoosuc up the gully contain knots of chalcopyrite.

Return to cars and continue up Route 123A.

15.1 Turn right down driveway of home of Dr. and Mrs. George Hanson, who will provide lunch for us. Continue past house and park in designated area.

STOP 5. View of the Alstead gneiss dome. The dome cover has been eroded through to the Oliverian gneiss core. Ammonoosuc volcanics, the Quimby Formation, and the Littleton Formation dip away from the axis of the structure.

Return to cars and turn right on Route 123A.

16.4 Turn right and cross one-way bridge over Cold River.

17.0 STOP 6. Park in lot of the Acworth town garage parking area. Cross road and ascend washed out mine road leading to the Beryl Mountain pegmatite quarry. This deposit was mined for beryl in the 1940's and for feldspar and quartz in the 1950's. It still yields some choice rose quartz and golden, white and green beryl. This is one of a number of pegmatites in the region, the so-called Keene pegmatite district. Climb crude trail on the left side of quarry to summit which is composed of the glacially polished quartz core of the pegmatite. Follow trail away from quarry face for good view of Cold River valley, the Alstead dome, and glacial lake terrace below. Note that pegmatite forms an isolated bump on the upper slope of Cold River valley.

Return to cars and continue along Beryl Mountain road.

18.0 Turn left at mineral shop.

22.5 Turn left on Route 123.

23.6 STOP 7. Kinsman porphyritic granite. This thin, elongate sheet-like pluton is occupied by the headwaters of the Ashuelot River. The Kinsman also forms a small knob on the left in which the not-so-famous Marlow Profile may be found with some difficulty. It resembles Richard Nixon or Fred Flintstone more than the stern features of thefamous New Hampshire rock profile.

Return to cars and continue along Route 123.

24.3 Turn right on Routes 10 and 123. PC-MAC Connection sells discount software for those computers.

25.1 Turn left on Route 123.

28.3 Turn into Pitcher Mountain parking lot on left. Secure cars and climb about 1/3 mile to the summit near look-out tower.

STOP 8. This elongate ridge is near the eastern edge of the BHA and is underlain by rocks that have been assigned to the Silurian Rangeley Formation. The first beds that we cross display rusty weathering and may be correlated with the Rangeley "C" in northwestern Maine. On either side of Pitcher Mountain, the valleys are underlain by the Kinsman granite. Note especially the chain of ponds in the valley immediately to the east. Further to the east lies the western limit of the Kearsarge-Central Maine Synclinorium, underlain extensively by the Littleton Formation. Mt. Monadnock to the southeast is part of the Littleton trend. Mountains composed of similar rocks in Massachuetts are hidden by Monadnock but are visible from the road ahead of where we parked. One is supposed to be able to see Mt. Greylock in western Massachusetts and some of the trap ridges in the Mesozic basin from here. To the north, Sunapee Mountain is underlain by Littleton rocks, with Rangeley units to the east. Mt. Kearsarge may be visible to the northeast. To the west, the Alstead dome forms a long ridge. To the northwest, in front of the Acworth church is Beryl Mountain and Ascutney Mountain beyond. Ascutney is composed of syenite that is related to Mesozoic hotspot activity. To the west lies the Connecticut Valley and the Green Mountains beyond.

End of trip. Those continuing east and to the Boston area continue on Route 123 to Route 9. East on 9 to I-89 and I-93 in Concord. To go southwest or north, return west along Route 123 to Route 12. Access to I-91 near Bellows Falls. To return to Keene, follow Route 123 back to Route 10, then left on Route 10 to Keene.

Thank you. Drive carefully.

# MESOZOIC STRESS HISTORY OF THE UPPER CONNECTICUT VALLEY AT TURNERS FALLS, MASSACHUSETTS

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## BACKGROUND GEOLOGY FOR THE TRIP

The Mesozoic Deerfield Basin lies north of the Hartford Basin, separated from it by the Amherst Inlier. The most prominent topographic marker is the Jurassic Deerfield Basalt (fig. 1), a flow unit separating the Triassic Sugarloaf Arkose on the west from the Jurassic Turners Falls Sandstone on the east. Filling of the basin was marked by eastward stepping of the depocenter from Triassic to Jurassic time to allow accumulation on Paleozoic basement of the Mount Toby Conglomerate along the Eastern Border Fault.

The Eastern Border Fault, about 6 km east of Turners Falls, is an excellent example of tectonic heredity. It follows more or less faithfully the trace of the west-dipping flanks of the Devonian Pelham Dome to produce the broad arc evident in figure 1. The dip of the fault is also subparallel to the dome units, about 30 degrees to the west. As mapped by Willard (1952), a series of smaller normal faults in the basin trend parallel to the border fault. Two of these faults will be examined on the trip, the Falls River and the Canada Hill Faults (fig. 2). Details of the proposed interpretation of the fault zones differ from those of Willard who extends them for great distances rather than splaying and dissipating the displacement into minor fault blocks and local folds.

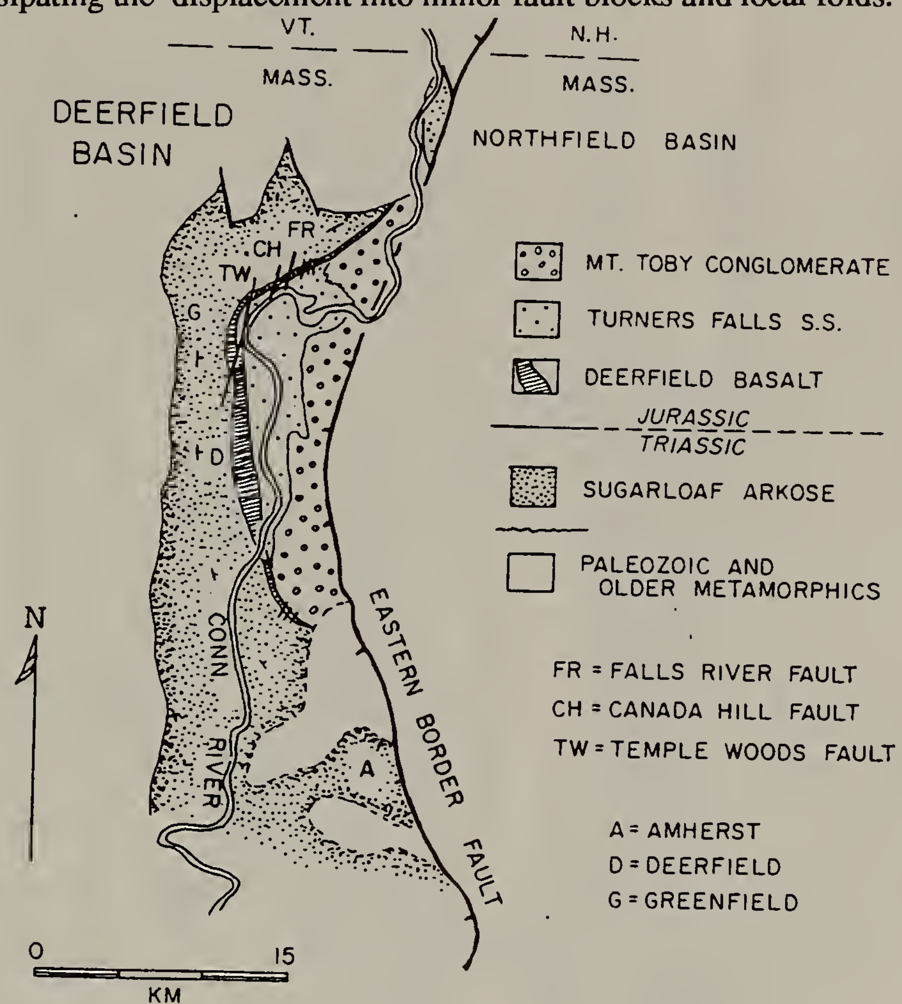



FIGURE 1. INDEX MAP FOR THE DEERFIELD BASIN

The stratigraphic summary of the field trip area is presented in figure 3 with lake bed zones being the primary means of breaking the column into the informal units suggested here. At times of low water there is over a kilometer of nearly continuous rock exposure just downstream of the Turners Falls Dam across the Connecticut River. As pieced together in figures 2 and 3, this repesents slightly more than 300 meters of Jurassic stratigraphy. The units are rich in sedimentary and igneous structures. (Members of the trip are encouraged to point out any features of interest to the group.)

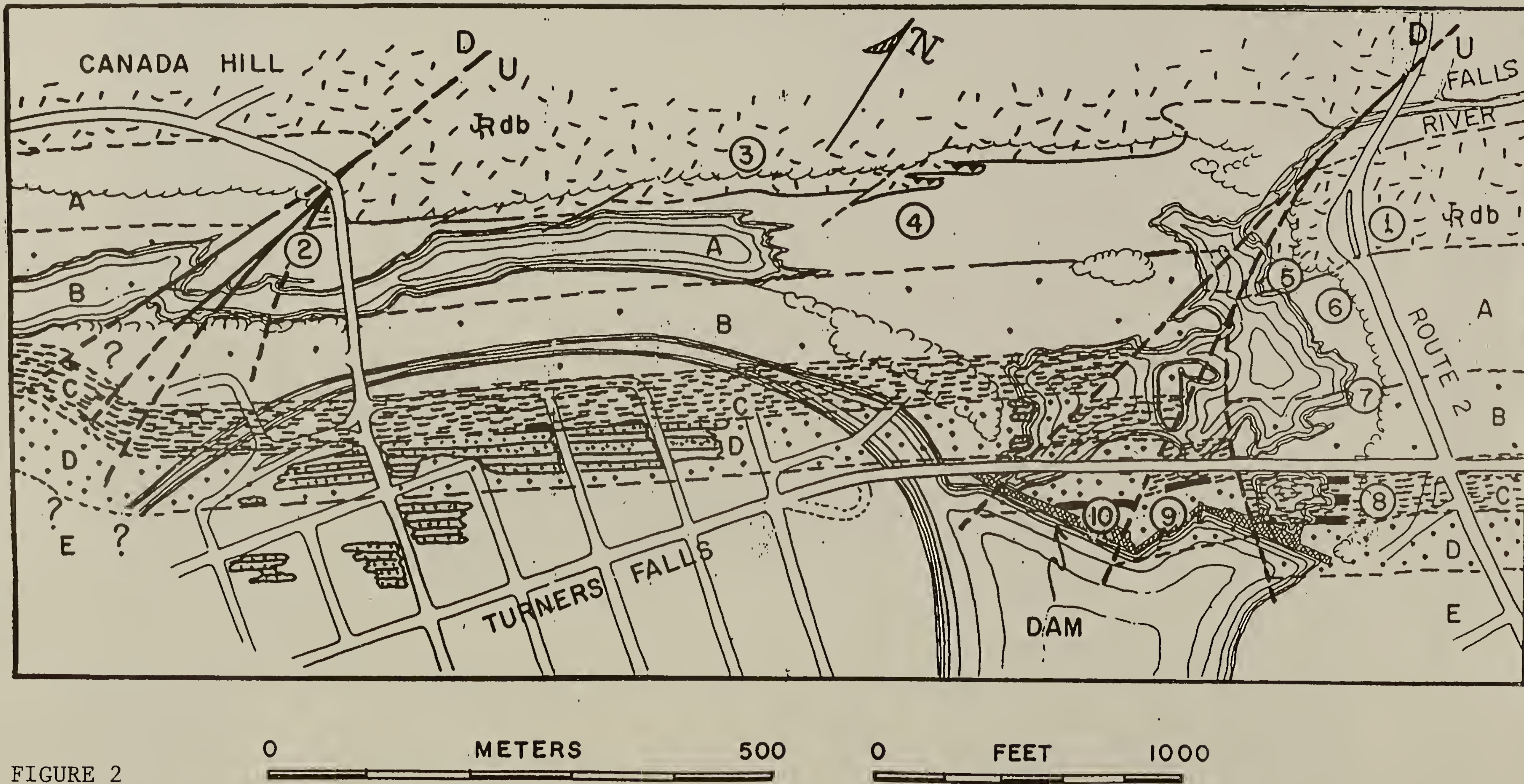
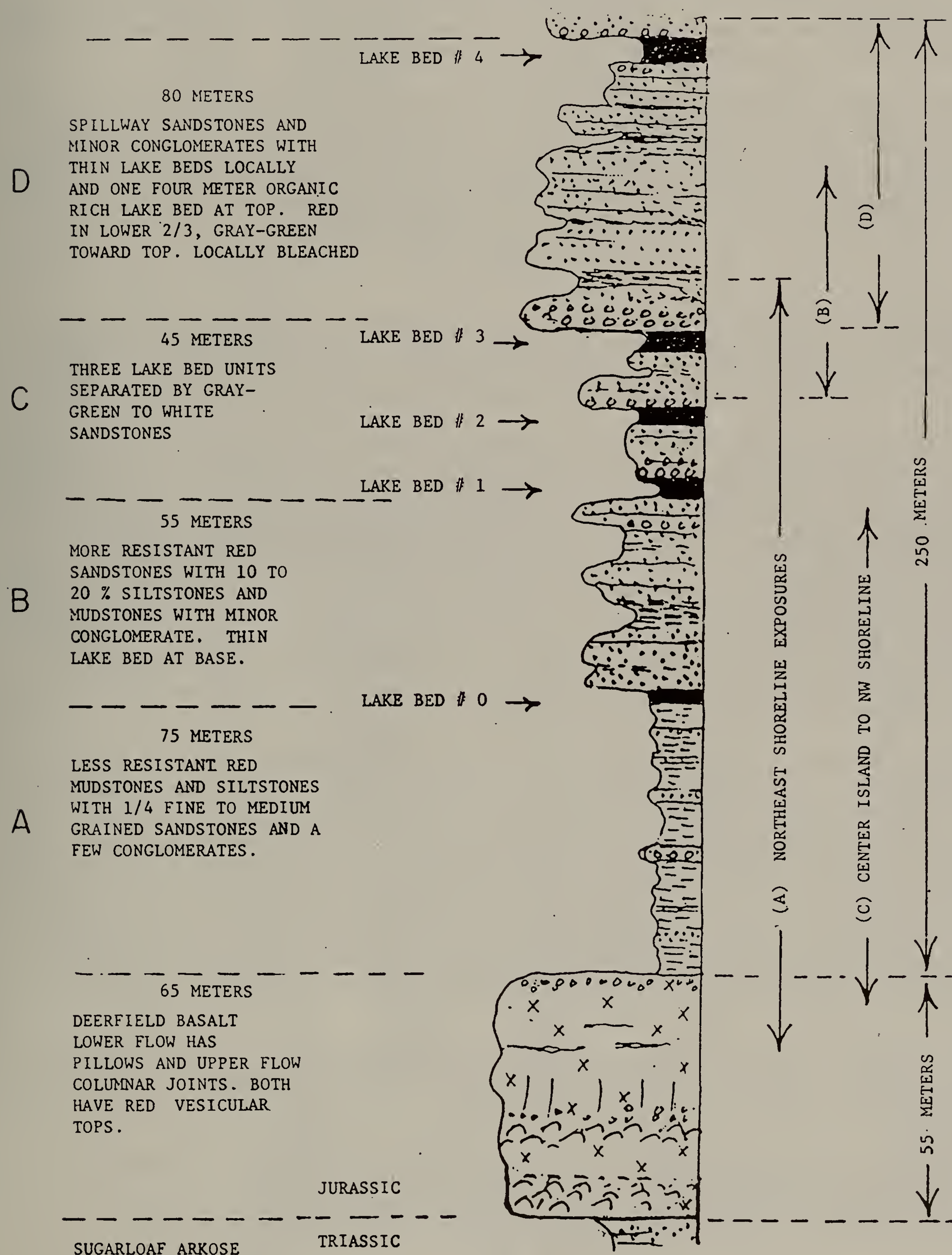



FIGURE 2

GENERALIZED GEOLOGY OF THE TURNERS FALLS AREA

FIGURE 3

# COMPOSITE STRATIGRAPHIC COLUMN TURNERS FALLS DAM, MASS.



The exposures (fig. 4) show a far more complex series of stuctures than those usually ascribed to the Mesozoic of this region. Such structues include significant normal faulting with NW-SE extension followed by conjugate strike-slip faults indicative of NE-SW compression. Local thrusting and small scale compressional folding are in part younger than the normal fault episode. Several periods of jointing and veining also were associated with these events, especially with the normal faulting and extension.

The 55 meters of exposed Deerfield Basalt stratigraphy include two lava flows, pillows, columnar joints, volatile-rich sill-like overpressured zones, neptunian dikes, and sedimentary interbeds folded in part by moving lava. The 250 meters of exposed Turners Falls stratigraphy reveal local isoclinal folds, extension breccias of several ages, and bedding plane faults decorated with extension growths of mineral fibers.

The rocks include a rich record of clues to changing fluid compositons, temperatures, and differential degrees of ductility during deformation. Cooling basalts with fluid and jointing effects, minor ore mineralization along fault zones, and oxidation-deoxidation effects in the redbeds and basalt are common in some areas. The organic-rich lake beds passed through the hydrocarbon maturation window during the major periods of normal and strike-slip motion leaving their marks on some of the veins and faults. Sandstones and conglomerates were lithified early while some of the fluid-rich lake beds remained ductile during most of the deformation.

## THEMES

There are two basic themes on this trip. One is to show a wide range of sedimentary and igneous features which gradually merge with time into tectonic structures which include extension, compression, and strike-slip motion under varying degrees of ductility and fluid environments. Also evident are the changing paleo-vertical versus present-day vertical effects on structures produced as the basin tilted to the southeast.

The other fundamental aim of this trip is to demonstrate that two major orientrations of Mesozoic stress (fig. 5) are recorded in this area as well as many minor and/or local stress effects and orientations. The **first major stress was extension with sigma 3 oriented approximately N40W.** The plunge of this sigma 3 as recorded in the rocks changed with time because the main tilting of the basin to the SE was associated with this phase of faulting. Orientation of the early sigma 3 is shown nearby in orientations of basaltic dike swarms and locally by similar orientation of many smaller normal faults, joint and vein orientations. The early fracture pattern is complicated by an oblique average orientation of the main faults in the dam area at N15E, possibly controlled by parallelism to the nearby Border Fault. Dips of these faults are relatively shallow to the NW, possibly synthetic with the Border Fault and in part a reflection of synchronous and subsequent tilting. The fact that overall sigma 3 is oriented about 20 degrees clockwise from perpendicular to this large-scale fault anisotropy, requires some oblique, right-lateral normal motion. (Portions of the Border Fault show similar effects as shown by Maher's study of drill cores included in Wise, 1979, p. 234-238).

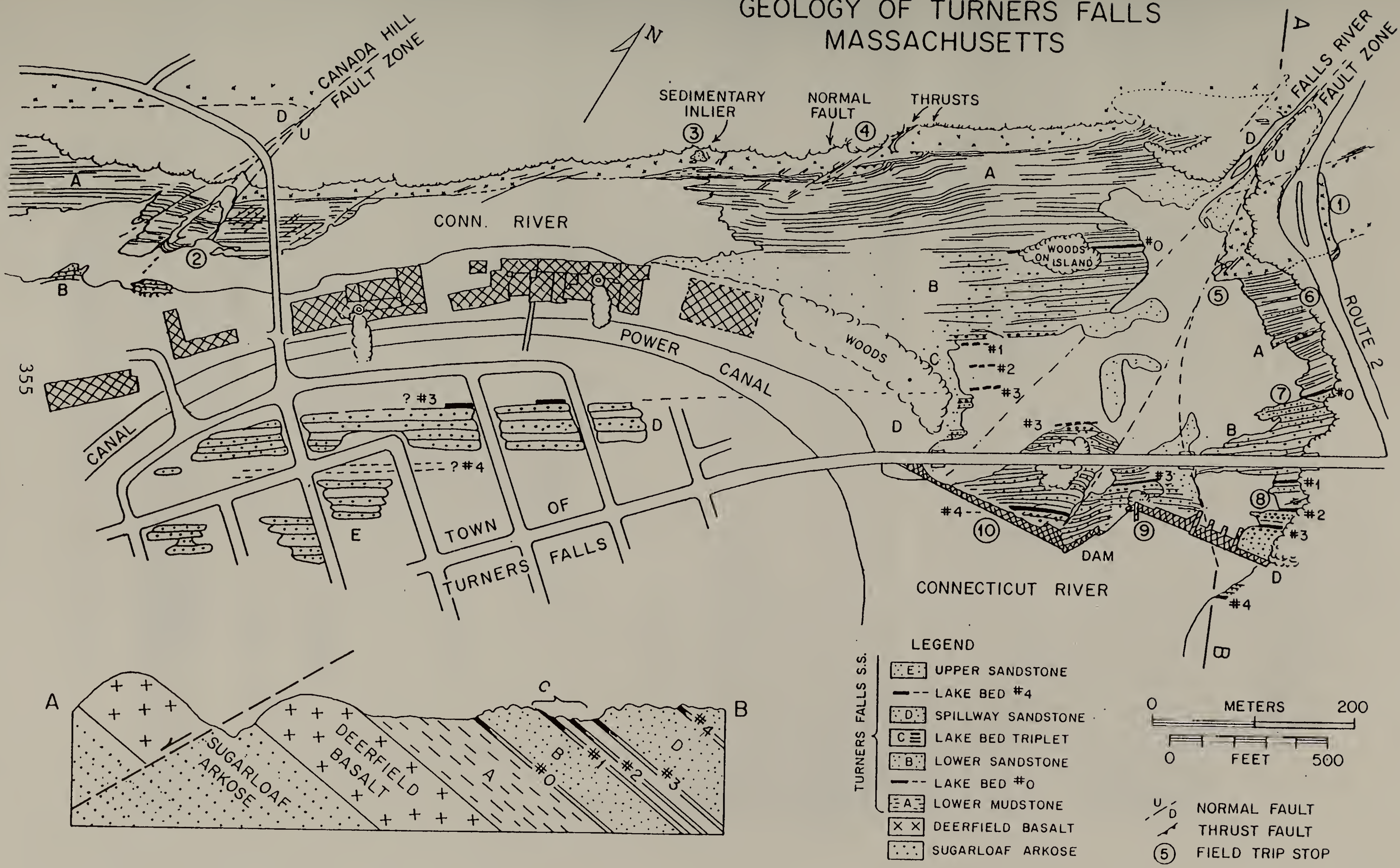
The **second major stress field involved compression and strike-slip faulting with sigma 1 nearly horizontal at about N40E** in the field trip area. Sigma 3 was relatively little changed from the first or normal fault phase, raising the question of whether sigma 3 was the time-constant major deviatoric stress with sigma 1 and 2 merely exchanging places from first to second stages. Effects of this second stress field were felt broadly as first pointed out by Goldstein (1975) in showing that strike-slip faults indicative of a NE to NNE sigma 1 are common in the northern part of the Deerfield Basin. (Actual sigma 1 orientation could be more northeasterly than interpreted by Goldstein because of possible inclusion of phase 1 oblique-slip motions in his data.) Further, Williams (1979) showed that many late-stage kink bands in the Paleozoic metasediments at the north end of the basin (fig. 5) also indicate a compressional phase with sigma 1 oriented N55E with a plunge of 25 NE. Precise orientations of this second phase of northeast compressional effects may vary locally but are certainly real and widespread in this region.

The overall fault and stress oreintations are shown in the cumulative plots of figure 6. For convenience of reference the faults are grouped by slickenside orientation clusters in figure 7 with a letter for each. Subgroups are based on differences in orientations of fault planes or motion senses.
The groups are:

A: slickenlines plunging NNW. Subgroups include normal and reverse as well as right-lateral faults;
B: slickenlines are horizontal and NE to E with dominant left-lateral motion;
C: slickenlines plunge shallowly S to SW and are dominantly right-lateral;
F: (location #10) slickenlines are on bedding plane faults marked by fibrous calcite growths;

FIGURE 4.

# GEOLOGY OF TURNERS FALLS MASSACHUSETTS



J: (location #5) slickenlines are on flat planes with NNE motion;
I: (location #3) slickenlines cut sedimentary inlier in basalt and plunge shallowly SSE;
P: (location #4) slickenlines are thrusts on flat, polished, hematitic surfaces with shallow plunges to the SSE;
V: (location #1) slickenlines are fairly close to the present day vertical.

The summary plot of poles to faults shows most faults strike E to NE to E and dip to the NW. Their average orientation is crudely perpendicular to bedding with dips varying up to 30 degrees on either side. A second set of fault planes has motion crudely parallel with bedding. The summary plot of poles to all veins (mostly carbonate and quartz) shows generally NW-SE extension. Some of the veins are crudely normal to bedding but others, unlike the faults, tend to have steeper dips to the NW than expected for perpendicular to bedding. The relationships are interpreted as some veins forming early in the extension process prior to significant tilting whereas others postdate much of the tilting of the basin fill to the SE.

In the Turners Falls area, the strike-slip, conjugate faults (Groups B and C, figure 7) post-date most of the tilting of the basin fill with all the strength anisotropies associated with differential rates of lithification. The result is a complicated geometry with some fault planes changing orientations abruptly from near bedding-parallel to near bedding perpendicular and back again. Associated local stress reorientations produce curving slickenlines and relationships which might be incorrectly interpreted as small normal or thrust faults if the larger fault pattern were not exposed. Best exposures of this complex, younger, strike-slip system are at location #10 but they are also present in reasonable prominence at locations #1, #7, and #8.

Flat faults, some of them thrusts and apparently unrelated to the strike-slip system, occur as separate deformational phases throughout the field trip area (Groups I, J, F, P). Some are indicative of NNW to NW to W compression and are most readily seen in association with the folding and shortening at location #4. Other locations are at #1 and #5. Some members of this group post-date the two main phases of deformation but I am not yet certain that all of them are so young nor that all have the same sense of displacement.

From a broader tectonic view, this trip is designed to demonstrate the older Mesozoic stress orientations as well as the younger reorientation of the stress field as shown in figure 5. Correlation of this fundamental change in the stress history can probably be correlated with the change from rifting to drifting as the Atlantic opened (De Boer and others, 1988).

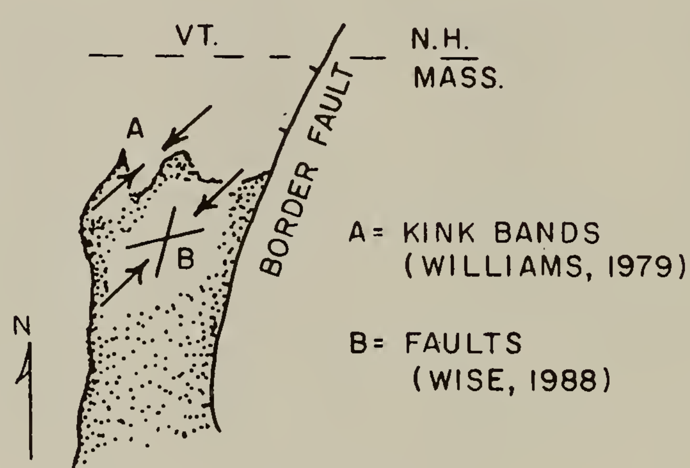



FIGURE 5. YOUNGER MESOZOIC STRESS UPPER DEERFIELD BASIN, MASS.

## CAUTION AND ACCESS

In spite of the short distance and relatively level topography of this trip, walking can be difficult. Some of the travel will be across one meter microrelief formed by tilted edges of thin sedimentary beds. Thin soled shoes can be very uncomfortable on such terrain. In case of rain and slippery rocks, the itinerary may be modified or even

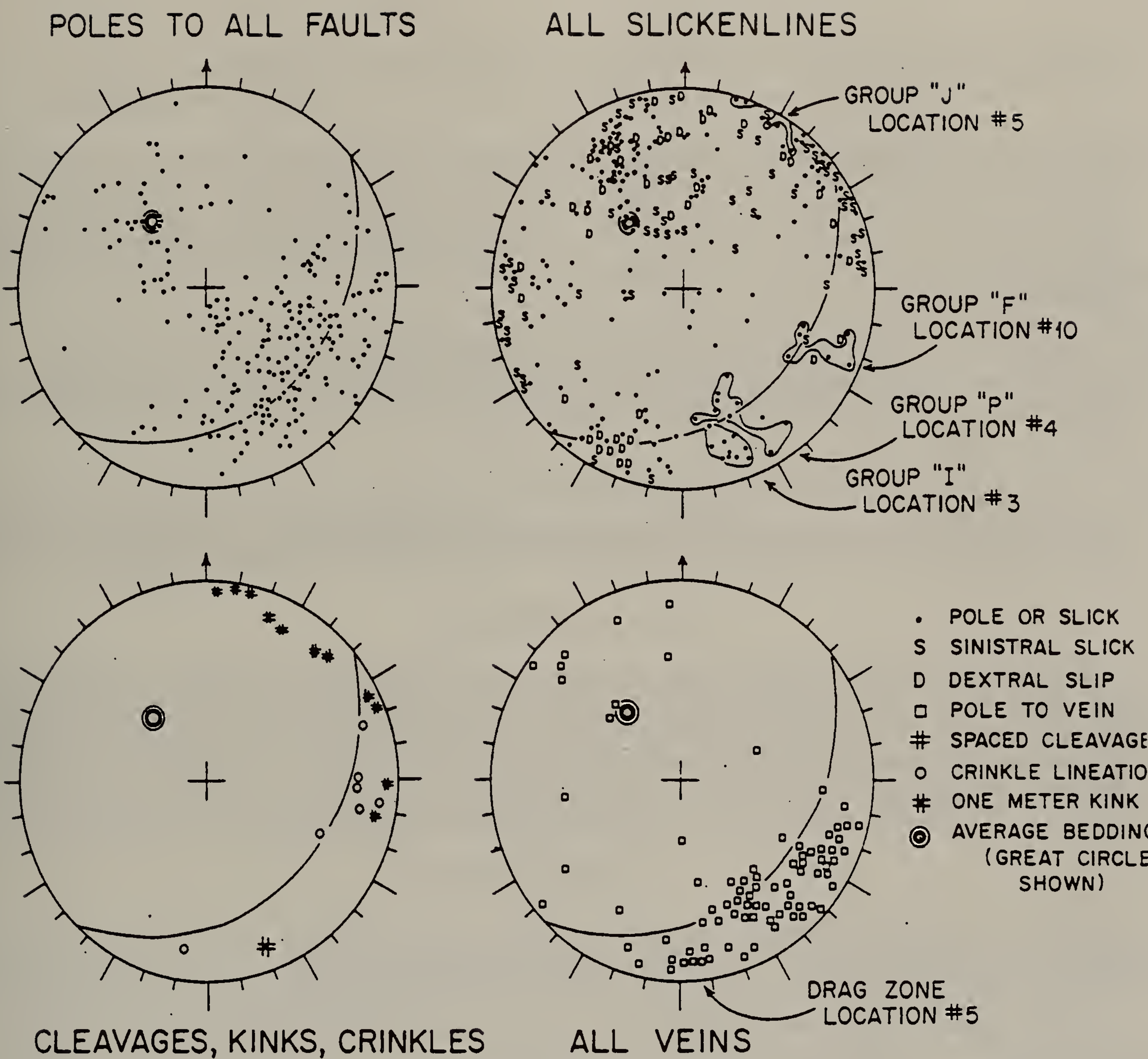



FIGURE 6. CUMULATIVE PLOTS OF ALL BRITTLE FRACTURE DATA AT TURNERS FALLS

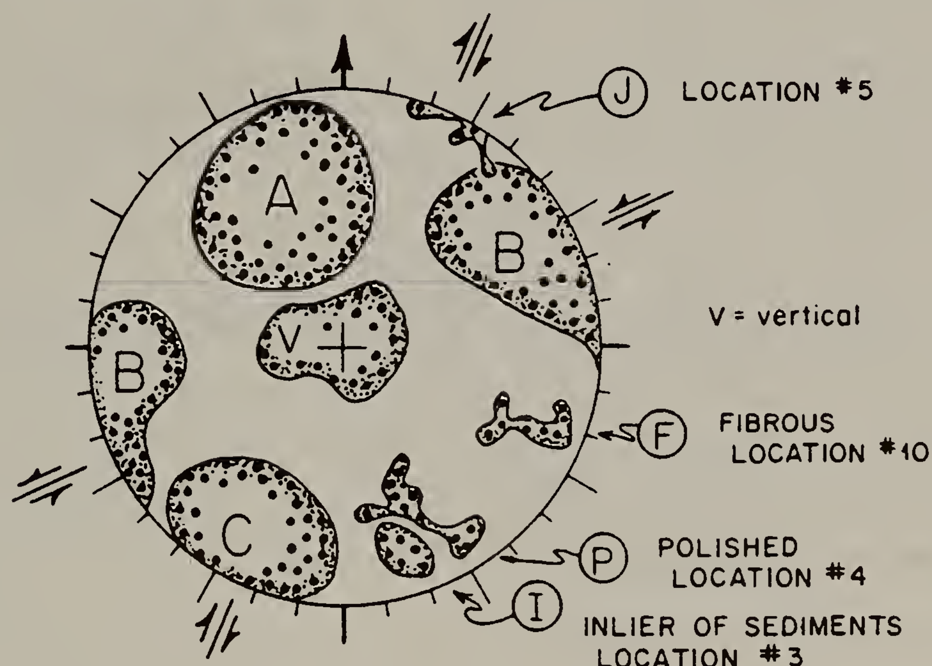



FIGURE 7. GROUPING OF TURNERS FALLS FAULTS BASED ON SLICKENLINE CLUSTERS

canceled for reasons of safety. For easier access to the spillway island, we are attempting to get permission from Massachusetts Electric Power Company at Northfield Mountain to use their walkway across the dam.

For anyone planning to run this trip at another time, remember that the exposures are downstream from a major dam. Most years, most of the exposures are accessible from June through September. In dry years this period can extend from mid-May to mid-November. Portions of the exposures along the northeast shoreline can be visited at almost any time other than flood stage or winter ice. During high flow periods, care should be exercised for changing water levels asssociated with opening and closing of flood gates. Slime covered rocks can make walking especially hazardous.

The road cut by the parking lot on Route 2 is always accessable and has a reasonably wide apron. However, because of high speed traffic on the blind curve, crossing from the parking lot to reach the cut can be dangerous. Use extreme care: a flag person or a lookout on the parking lot side of the road is recommended for the return transit of any large group.

Outcrop locations #3 and #4 can be reached from the Route 2 parking lot without fording Falls River. Cross westward over the Falls River bridge and immediately go down over the embankment to the south. A footpath leads over the basalt cliff about 10 meters above water level to exit onto the Connecticut River floor.

## ITINERARY

The trip will convene on Sunday, October 16, at 9:00 A.M. at the Route 2 parking lot overlook of Turners Falls Dam, just north of Turners Falls, Mass. First we will look at the road cut opposite the parking lot (Location #1, figures 2 or 4). Most of the cars will be left at this overlook as the group is ferried to the next bridge downstream. (Parking there is possible at the southeast end of the bridge next to the Franklin County Housing Authority). The group will cross the bridge and descend to the river floor southwest of the bridge abutment. We will follow the shoreline back to the parking lot by lunch and then examine the spillway area after lunch.

## A BIT OF HISTORY

Prior to white occupation, this area was the home of the Pocumtuck Indians, a relatively peaceful group of the Algonquin peoples. The large falls of the Connecticut River constituted an ideal spot for fishing and the adjacent terraces provided good land for farming. Many artifacts have been found in the area.

Adrian Block first sailed up the lower Connecticut River in 1614 and by the 1630's William Pyncheon had established a navigation system along the river with towns beginning to be established at the expense of Indian lands. Farther south in Connecticut some of the newly founded towns were attacked by Indians. The settlers retaliated by massacring 600 Pequot Indians near Mystic in 1637. However, in this area the Pocumtucks had relatively good relations with the settlers and even sent canoes loaded with corn down river in 1638 to help some of the towns impoverished by the Pequot War. Later, the Pocumtucks had a misunderstanding with the Mohawks. The result was destruction of much of the tribe in 1664.

Finally, in 1675 an Indian uprising known as King Phillip's War brought local feelings to a boil. In late spring, 1676, a group of about 300 Pocumtucks were camped on the terrace at the north end of the present Turners Falls bridge. Subsequently, about 80 head of cattle disappeared from the area just to the southwest. To avenge this, a band of 142 settlers led by a **Captain William Turner** of Hatfield rode to the west side of the basalt ridge of the field trip area. Shortly before dawn on May 17, 1676, this group left their horses just west of the ridge to sneak across Falls River and over the ridge near the Location #1 parking lot. The Indians were all asleep from a feast the previous night and had no sentries posted. The attack killed most of the 300 inhabitants of the camp, a few trying to escape over the falls by canoe or swiming. Only one settler was lost.

The victorious party headed south very quickly in poor array, possibly because of a rumor of the approach of King Phillip and 1000 of his warriors. Captain Turner foolishly allowed the force to break into several small groups which were attacked separately within a few miles by the Indian survivors and by Indians from surrounding villages. About 40 of the settlers were killed in the retreat, including Captain Turner as he was crossing the Green River in present-day Greenfield. His second in command, a Captain Samuel Holyoke of Springfield, finally established order and moved the survivors back to the safety of the stockade at Hatfield.

A tombstone-like monument at the north end of the bridge marks the site of the massacre and the events whereby Turners Falls got its present name.

More Recent History

1798 Canal and locks built around the falls. Authorized by Massachusett legislature and signed by Govenor John Hancock.
1826 First steamboat passes up the canal
1856 Canal and locks abandoned as railroads take over.
1868 Establishment of Turners Falls. (No real town existed before.) Alvah Crocker, after building the Fitchburg Railroad, laid out the town as an industrial community and convinced major industry to establish there. This was one of the last of the large canal-based water powered industrial communities to be established in the region.1872 Town of Greenfield sets up competetion by establishing Falls River Industries along Falls River, 1/4 mile west of parking lot stop. Stopped when Montague Paper Co. buys rights to the water of Falls River and lays a 24 inch pipeline across the river bed. (The trace is still visible as a trench just upstream of the wooded island in mid-stream.)
1879 Paintings show main factories completed, town more or less in its present state and two suspension bridges across river.
1950 - ? Industrial decline as tool industries close or move out. Shoreline factories are now owned by paper companies.

ACKNOWLEDGEMENTS

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FIGURE 8. ROUTE 2 ROADCUT IN DEERFIELD BASALT TURNERS FALLS, MASS.

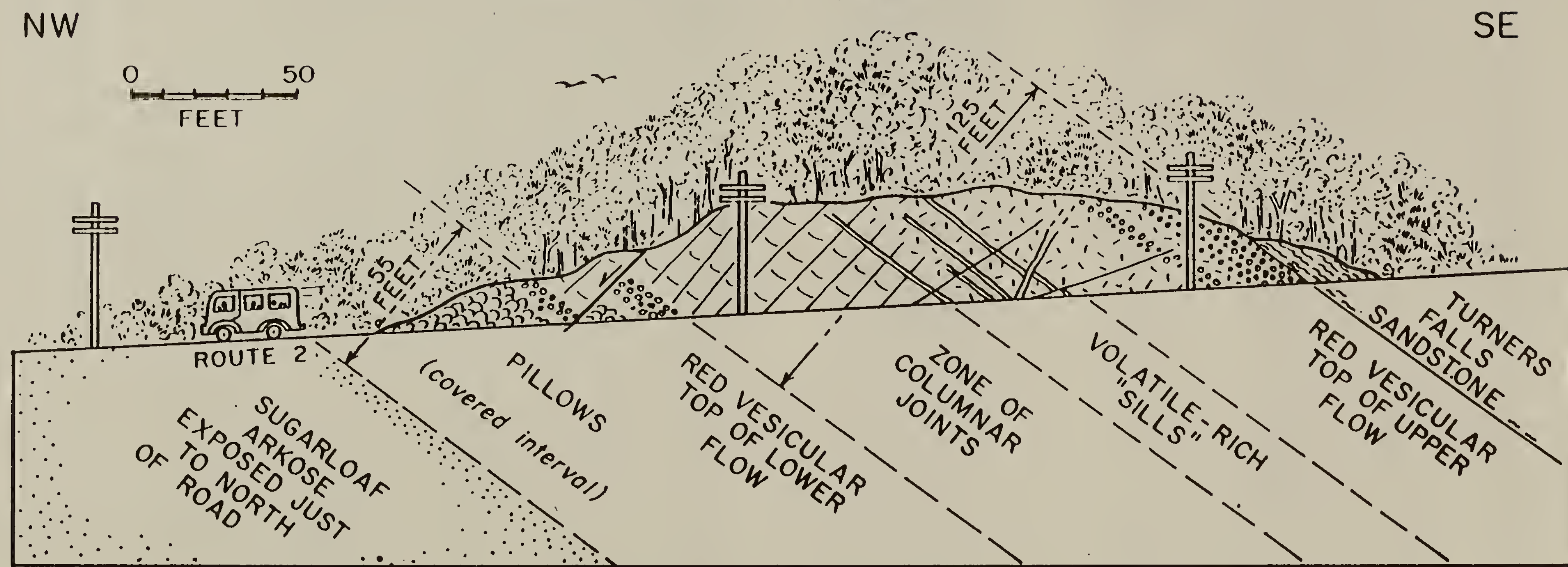
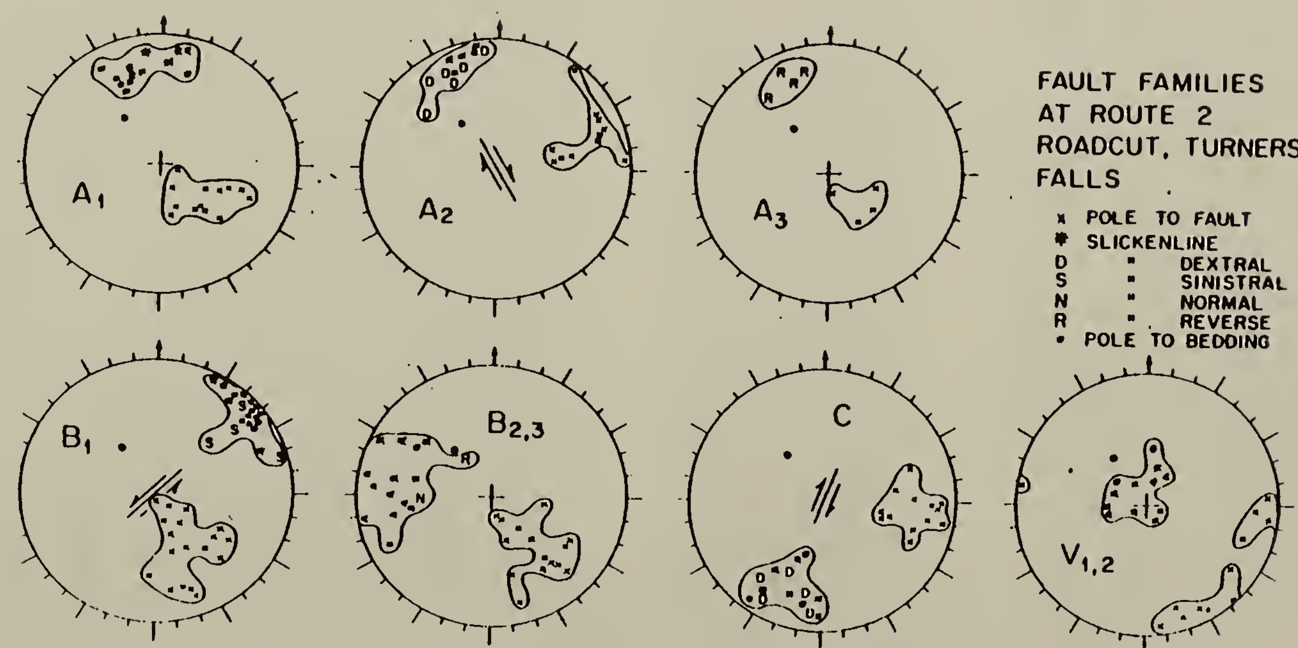



FIGURE 9.



## NOTES FOR SPECIFIC LOCATIONS

**LOCATION #1: Road cut opposite parking lot, Route 2**

-- Exposure of the Deerfield Basalt (fig. 8). Top of the flow is visible at right end of outcrop. Base can be found near the river in the woods just to the left. Good exposure of pillow lavas at the base can be seen in the next road cut across bridge to the west (another fault block) but use extreme care with traffic. Even better exposure is at the rear of the large sand pit just south of Route 2, 1/4 mile to the west.

-- Two separate flows. Stratigraphic zones are given in figure 8. Several sill-like zones 5-10 cm thick are marked by very coarse grain, diabasic texture, and local oxidation. These are interpreted as trapping of volatiles at different lithostatic pressures within the cooling mass. Alteration to near greenstone condition is common throughout flow. Note gradation of texture and grain size in upper flow, changing from fine at base, through medium, to very coarse grained and finally to vesicular at the top. Volatiles probably responsible for red oxidized tops of both flows.

-- Large columnar basalt joints in middle section show excellent arrest marks in paleo-horizontal orientation. Columns cut and hence postdate the sill-like, coarse grained zones. Note superimposed slickenlines with strike-slip motion approximating present day horizontal. Interpreted as result of strike-slip phase of faulting being largely post-tilting of basin.

-- Many tectonic joints, most about N45E, with assiociated calcite and some quartz. Some show pull-apart structures indicative of down to NW motion during filling. Interpreted as part of the general extension and tilting of the basin.

-- Fault with several meters of dip-slip displacement at left end of crop. Normal with relatively little mineralization but also includes some strike-slip motion. Orientation N30E.

-- Complex families of minor slickenside motions are separated and designated by letter in figure 9. Displacements may be only a few mm but their strain patterns give some indication of the causal stresses. In the absence of displaced markers, motion senses are best determined by transpressional and transtensional effects associated with irregularities on fault surfaces. Mineralization is concentrated in the low pressure zones in the lee of an asperity whereas better polish and grooving are on the transpressional side. (Some of the best examples are on the strike-slip slickenlines crossing the arrest marks on the columnar joints.)

A number of fault families are present as illustrated on figure 9. Most represent reactivation of pre-existing joint surfaces. The typical phase 1 normal faults are the A1 group (subsequently tilted to the present shallow dips). The A2 group has identical slickenlines, steep dips to the west, and right-lateral motion; it is interpreted as the result of oblique strain produced by the N15 E Border Fault and Falls River Fault interacting with the average regional NW-SE extension direction. The A3 group is a younger reversal of motion on these older fracture planes. B1 and C are the left and right-lateral strike-slip fault groups associated with the younger NE-SW sigma 1 orientation. Groups B2 and B3 represent shallow to moderately west-plunging, right and left-lateral motions on NW dipping planes. Some of the reverse and right-lateral oblique motions may be part of the B1 conjugate system. The normal, left-lateral, oblique motions may be variations on the older extension pattern, a more extreme variation that group A4 at location #4. The V group of faults with slickenlines approximately vertical, may be partly a conjugate pair to the A1 extension faults subsequently tilted to the present vertical from a 60 degree SE dip and plunge.

-- Mineralization on the A set of faults is largely calcite with some quartz. On the B and C sets, minerals are similar but more sparse. The B3 or reverse set is marked largely by polished surfaces and hematitic stains.

-- Cross-cutting or overprinting relations of slicks are not common but locally show the C2 set to be young. A large overhang near the right hand end of the crop has three distinct sets of slicks: oldest is N45W and normal; next is N80W and normal; youngest slicks are N60E, top to the NE and hematitic stained, possibly an aberant member of the C set.

-- In general, note that strike-slip motions are younger than dip-slip ones wherevever age relationships are visible in the crop. Further, note that the strike-slip slicks are more closely related to the present-day horizontal than to paleo-horizontal.

FIGURE 10.

**LOCATION #2: Canada Hill (Normal) Fault Zone**
-- Drive through town of Turners Falls to bridge at downstream end of area. Park beside Franklin County Housing Authority and walk to west end of bridge. Descend to river bed just beyond the southwest corner of bridge. While crossing bridge, note the strong development of N15E joints parallel to the fault. Details of the air photo (fig. 10) can be appreciated from the bridge elevation.

-- Purpose of the stop is to:

(1) See the style of fault splaying into separate blocks as normal displacement passes from a strong and very brittle unit (the basalt) into a relatively weak and slightly less brittle sedimentary cover;
(2) See the pattern, typical of this area, of localization of N15E, 40 NW joints parallel to normal faults in their upthrown block; and
(3) See a half size, better exposed version of the Falls River Fault which dominates much of the rest of the trip.

-- The top of the Deefield Basalt is displaced horizontally by about 100 meters at this location. (The contact on the downthrown block is exposed in the riverbank a few hundred meters downstream and can be projected to the approximate displacements shown on the maps.) This displacement would correspond to about 120 meters of throw with respect to present vertical or, more probably, to **about 80 meters of throw** with respect to pre-tilt or paleo-vertical orientation.

-- Splaying of the fault upward from the basalt into lower units of the Turners Falls Sandstone, defines isolated and only slightly rotated fault blocks. The fault zones are not particularly wide nor intensely disturbed although local folding is visible. The exposes are all in the relatively monotonous beds of unit A and provide poor opportunity to determine precise displacement at any given location. Across the river to the southeast, shoreline exposures of resistant unit B show local disturbance on the projection of the splaying faults. Note dip of the normal faults in relation to present versus paleo-vertical.

-- Under bridge, note the dominance of N15E joints as spaced macrojoints extending for several tens of meters. Also note the relatively flat dip of the joint planes to the west as a reflection of pre-tilt geometry. The joints are interpreted as extensional structures localized by strains associated with the evolving Canada Hill Fault Zone.

**LOCATION #3: Sediments interlayered with top of the basalt**
-- Proceed upstream from location #2 for about 400 meters. If necessary use map and buildings across river to locate the spot on figure 4.

-- Two to five meters above normal water levels are interleaved slabs of fine grained sandstone in the upper 3-4 meters of basalt. Strike is generally N45E parallel with regional orientation.

-- These **sediments were localized in a low place in the top of the basalt,** possibly near the confluence of two flow channels. At extreme low water, the precise contact of the basalt can be traced as an interfingering contact with the sedimentary cover just below a lens of mudstone. This lens extends for about 100 meters and thickens from about 1 meter on either side to about 3 meters, across the former depression in the basalt topography.

-- The sedimentary slabs are **folded in two styles** (see figure 11).
(1) One style has steep to locally overturned dips, amplitudes of up to a meter, and basalt contacts which truncate and disrupt sediments on some of the steep limbs. These lines of truncation trend generally N 40 to 90 E. The truncated folds suggest overturning to the SE as being more common. The relationships are interpreted as final stages of Deefield Basalt flowage (here moving to the SE) interacting with the first channel fills of the Turners Falls Sandstone. Complex relationships also exist with the sediment at the base of the flow as Emerson (1898) gives lengthy descriptions of sediments being squeezed upward for many meters into what we would now term pillows lavas.
(2) The second fold style is more open with axes oriented about N75E, 10NE. The folds seem to be associated with the flat faults discussed below and have an orientation approximately perpendicular to the fault motions. Locally a crinkle lineation on bedding is parallel to the fold axes.

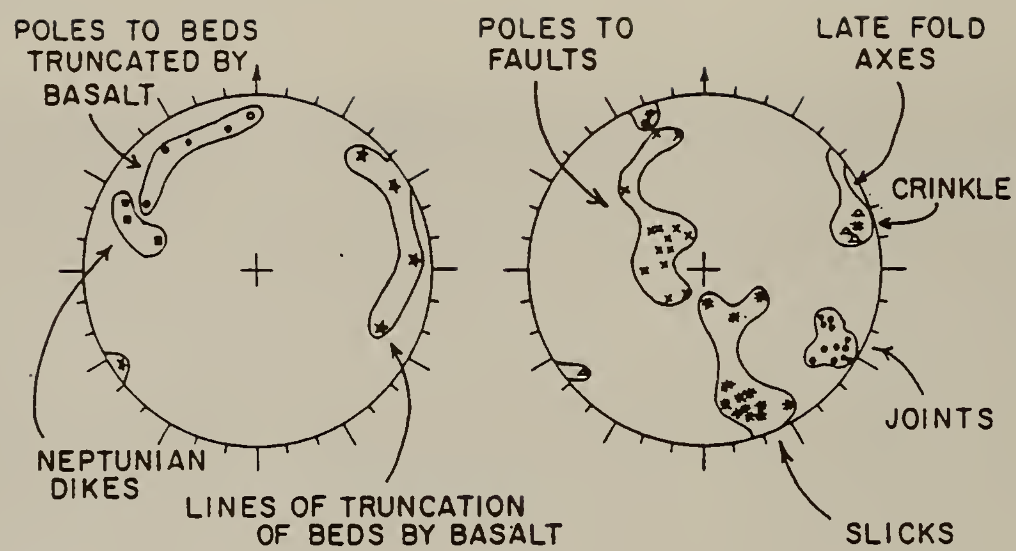



FIGURE 11. GEOMETRIC RELATIONSHIPS OF SEDIMENTS AND MOVING BASALT AT LOCATION #3

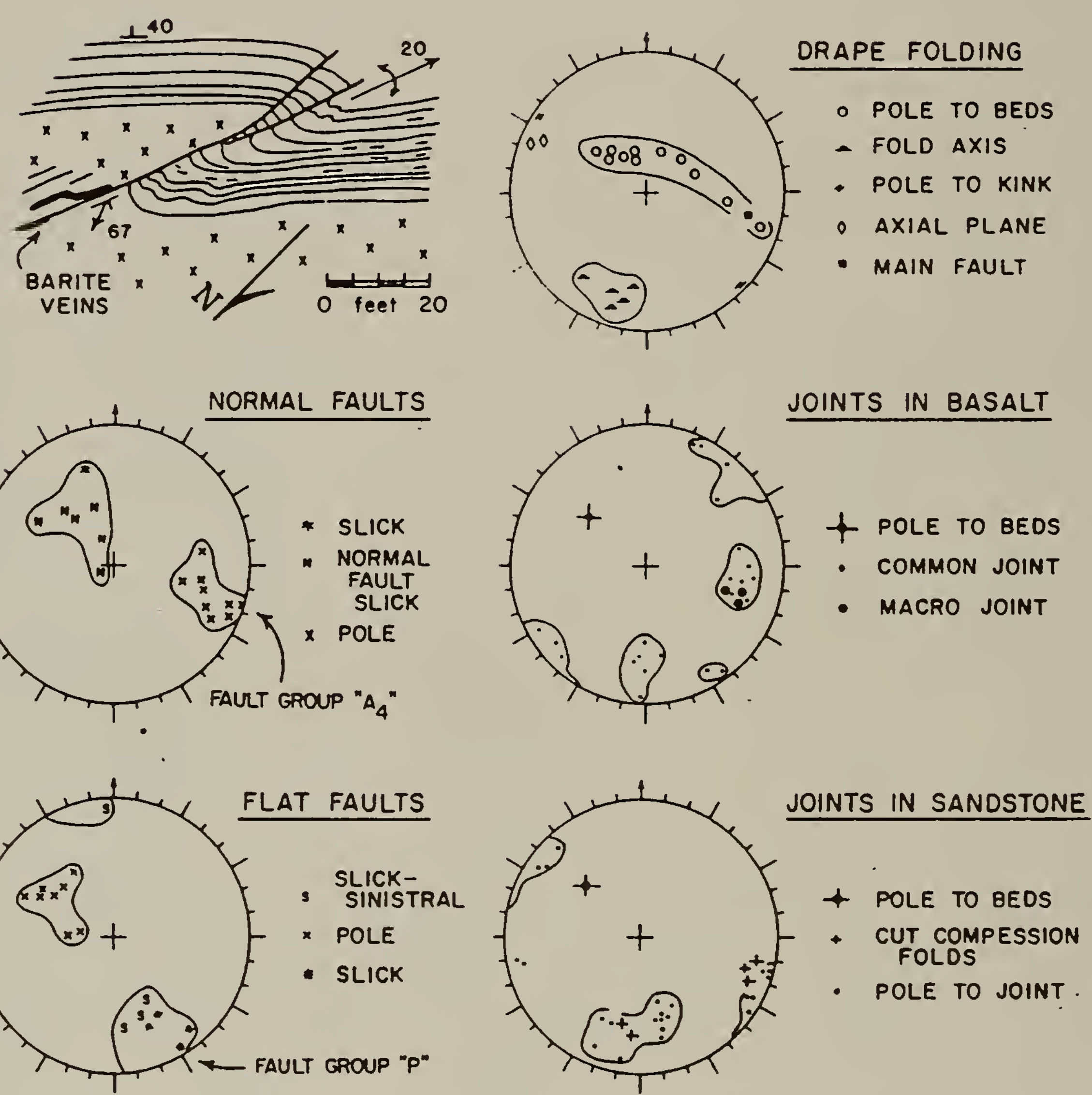



FIGURE 12. NORMAL FAULTING, DRAPE FOLDING AND JOINTING AT LOCATION #4

--A **pervasive set of minor fault surfaces** cut the sediments at dips of 10 to 30 degrees to the SE. These are largely unmineralized, grooved and curving surfaces with slicks plunging rather consistently at S20E at 20 degrees. The surfaces are relatively well-healed and are folded in sympathy with the enclosing beds by the second set of folds (above).

-- One well defined **joint set** at N40E, 70-90NW seems to cut the slicks and not be displaced by even minor fault motions. Some faint silicification follows the some of the joints. The joints curve slightly in passing through the folds.

-- The flat fault surfaces with slickenside orientations on this crop are similar to those in many other parts of the area. I am still not positive about the motion sense of these particular slicks, nor of the correlation of the joint set with other sets in the area, nor of all the relative age relationships in this outcrop. In light of this uncertainity, correlation of these structures to adjacent outcrops seems premature; help or advice would be appreciated.

**LOCATION #4: Outcrop sized normal and thrust faults**
--**Normal fault** (fig. 12) Is about a 1/15 scale Canada Hill Fault and 1/25 scale model of the Falls River Fault

+ Splays upwards out of basalt into sediments to produce slightly rotated fault blocks with some minor folding.

+ Other parallel minor faults cut the contact about 20 meters up and downstream from the larger fault.

+ Orientation N15E, 65 W ( approximately paleo-vertical. orientation.) Displacement of main fault of about 4 meters (with respect to paleo-vertical) dies upward in the next 25 meters of stratigraphy as a series of gently curving beds.

+ Tight folding, kinking, and crowding largely restricted to dragged synclinal area of downthrown block. Hinges are S15-25W, 20 SW.

+ Mineralization includes quartz, calcite, and barite (in masses up to 10 cm across) as well as minor sulfides. (Please refrain from mining the barite for lab collections.)

+ Prominent joint zone in basalt is parallel to the fault in the upthrown wall. This zone is approximately 5 meters wide with many younger minor fault motions and veins along individual joints. It is interpreted as an extensile feature of the brittle basalt as the rock was starting to flex in the early stages of fault development.

-- **Thrust faults**. Two examples are 50 and 100 meters upstream from the normal fault. (See figure 13).

+ Each fault zone offsets the basalt contact by about 3 meters. SW fault leaves a highly deformed one meter s eptum of sediments extending for about 20 meters in the basalt.

+ Farther up the bank (at the tree edge between the two thrust areas) is a one meter septum of sediment included in the basalt. Unlike the other septa at location #3, this does not show extensive interaction with moving basalt. Superficially it shows little evidence of metamorphism. Slickenlines can be found on its upper surface indicative of some movement along the zone. It is intrpreted as being some combination of interbedding and thrust repetition. The presence of this septum and possibly of others like it may be the cause of localization of more prominent thrusting of the basalt contact at this place.

+ Most obvious effect of the compression is to cause intense local crowding and folding of the overlying sediments. Fold hinges are oriented about N90E with plunges of 10-30E. Curvature of outcrops associated with these folds can be traced upward through about 40 meters of stratigraphy in the riverbed.

+ Good motion vectors on slickensided fault surfaces are rare on the thrust contacts. However, the area just to the SW has many flat, polished and hematitic, slickensided surfaces with motion vectors S20-30E and a plunge of 20 SE (fault group P on figure 12). Where determinable, the motion sense is top toward the NNW (thrust for present vertical orientations).

+ Relative ages??

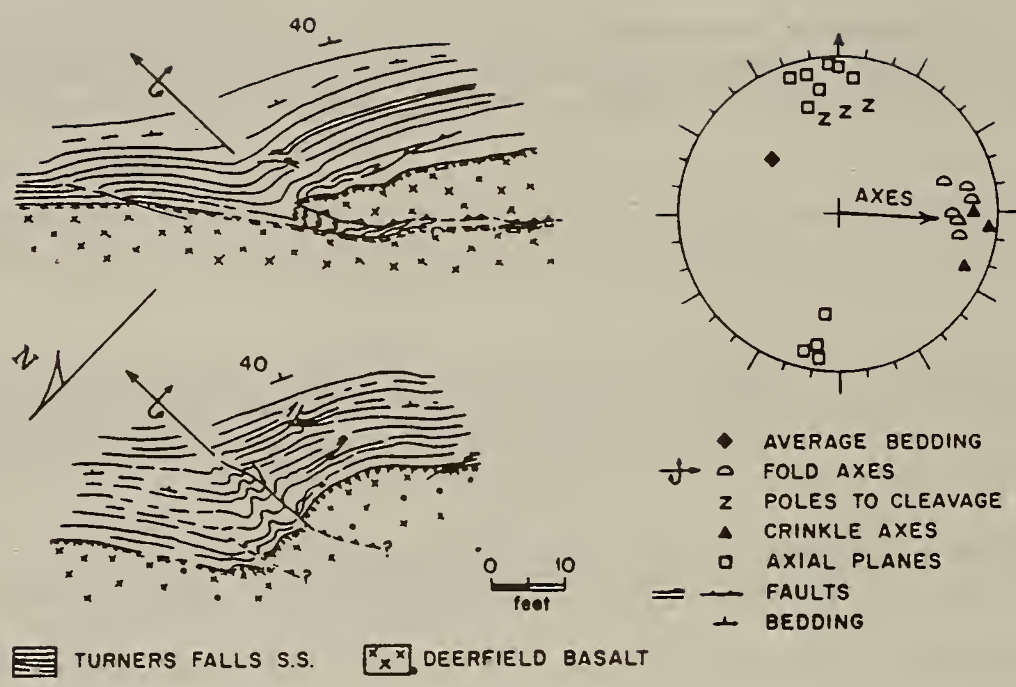



FIGURE 13. COMPRESSIONAL FOLDING AND THRUSTING
0.8 KM DOWNSTREAM FROM TURNERS FALLS DAM

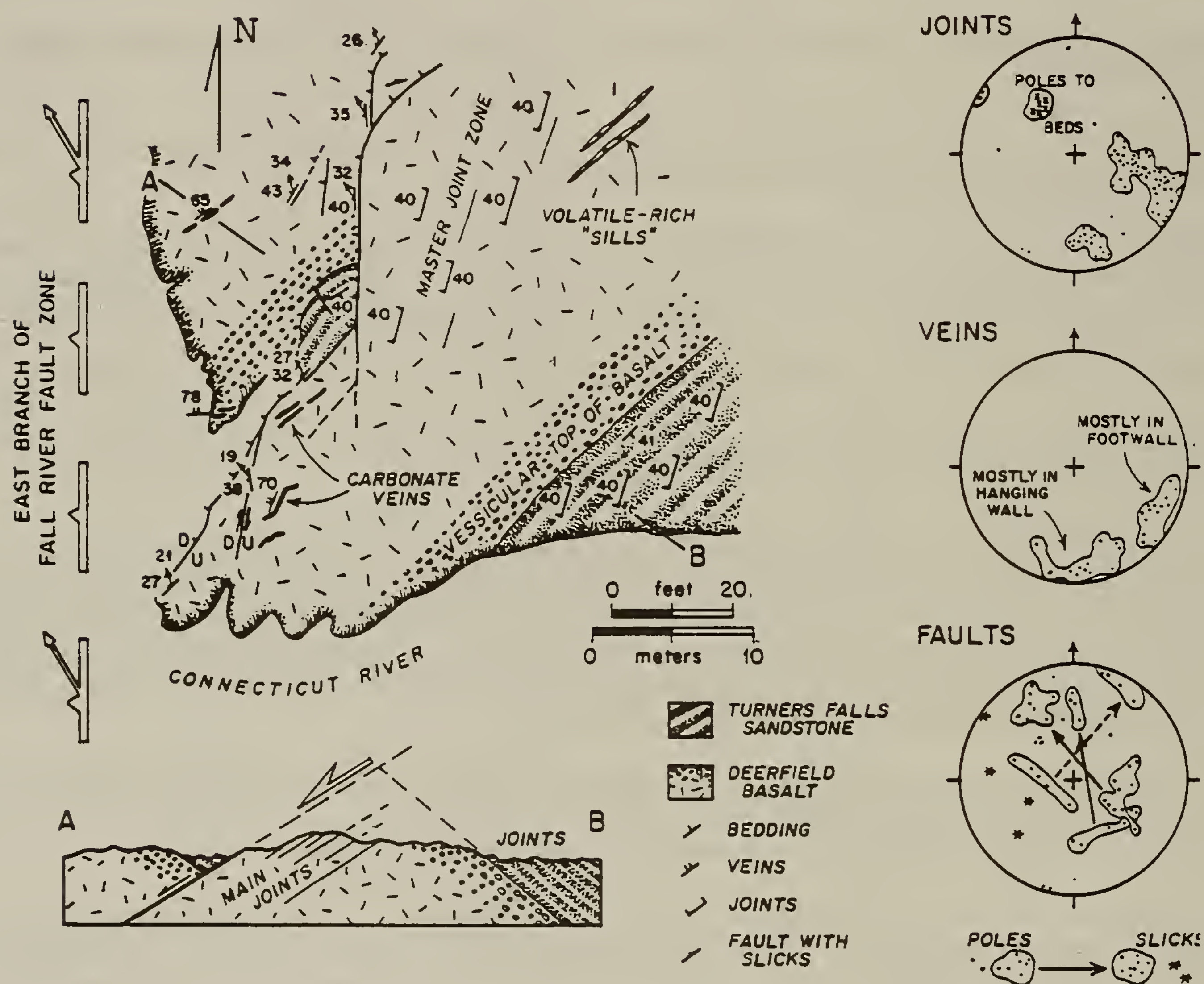



FIGURE 14. FALL RIVER FAULT DETAILS, CONNECTICUT RIVER EDGE
100 METERS SW OF ROUTE 2 PARKING LOT, TURNERS FALLS, MASS.

.. The flat slicks in the thrust area are similar in orientation and appearance to those in the sedimentary inlier at location #3.
.. The late folds in the inlier are sub-parallel with those in the thrust related area.
.. N30E joints are superimposed on the inlier faults and folds but N15E joints are superimposed on the sediments crumpled by the thrust faults. The N15E joint orientation is dominant in the adjacent normal fault zone.
.. One exposure has the hematitic, flat slickenlines clearly superimposed upon N15E joints of the disturbed zone of the normal fault.
.. The compressional folds associated with the thrusting, strike at an angle of only about 70 degrees to the flat, polished and hematitic slickenlines. It is possible that these structures are unrelated in time and stress.
.. The normal fault zone has associated hydrothermal mineralization whereas the compressional folds and faults are almost devoid of mineralization.

.. PRESENT WORKING HYPOTHESIS. There were several periods of N to NW to W directed thrusting, The one associated with the compressional folding postdated the period of normal faulting.

-- Depending on water stage, in walking upriver to location #5 we may ford Falls River or take the pathway up and over the basalt cliffs on its west side about 10 meters above water level and proceed across the Route 2 bridge to the parking lot. At the level of Falls River there is near exposure of the contact of sandstone upon basalt plus some minor faulting. The main Falls River Fault must pass up the river just east of these shoreline crops with a modest dip westward under the cliffs.

**LOCATION 5: River's edge down the path on the basalt from the Route 2 parking lot**
-- The upper contact of the basalt is traceable from the Route 2 road cut down over the valley slopes to the riverside exposures. Beds in the island on strike in the river are about 90 meters higher in the stratigraphy. Thus, the Falls River Fault must pass directly offshore from the basalt outcrop. A low water map of the end of the basalt ridge is given in figure 14.

-- A prominent joint set (upper net, right side of figure 14) marks the ridgeline used for the descending path. This class of joints at about N10E, 40W are most strongly developed in the basalts and overlying sediments adjacent to the normal fault zones.

-- Similar joints extend upwards into the covering sediments but show quite varied dips. Locally they can be seen with listric shapes opening toward the main fault. They are part of the extension process which ultimately produced the main fault.

-- A remnant of infaulted sediment is preserved and exposed at very low water level (fig. 14). Unlike the inliers at location #3, these sediments are clearly fault bounded with relatively shallow plunging slickenlines. The faults (map and lower net of figure 14) curve back and forth between the main west-dipping joint set and a moderately NW-dipping set of faults (fault groups A1 and A2). The parasitic nature of the faults is obvious as they make use alternately of the two main joint sets. The two groups of fault orientations are separated on figure 14 with an arrow pointing from the clusters of poles to the clusters of slicks. Between the two sets of parasitic (making use of older fractures) faults, the average motion on the fault must be at about N20W at a plunge of about 30 degrees.

The faulting involved little rotation of the displaced block because dips in the block are essentially the same as in the overlying sediments.

-- Sill-like coarsely crystalline zones, similar to the road cut, ar found at the base of the path. However, they can be seen to break upward irregularly across section at a few places.

-- The upper contact of the basalt can be followed away from the river to see some much poorer versions of neptunian dikes than are visible at location #3.

**LOCATION #6. CLASTIC DIKE IN THE TURNERS FALLS SANDSTONE**
-- Go upstream about 30 meters directly under the power lines high overhead. **Please refrain from picking apart the dike.** It's the best one around these parts.
-- Depending on recent river deposits, the dike may be seen (fig. 15) to have no connection with the beds above or below it at this level of exposure. A bedding- parallel septum in the middle of the dike may be a remnant of a sand bed downdip which was mobilized for the injection of the sand. The dike itself is segmented with mudstone septa

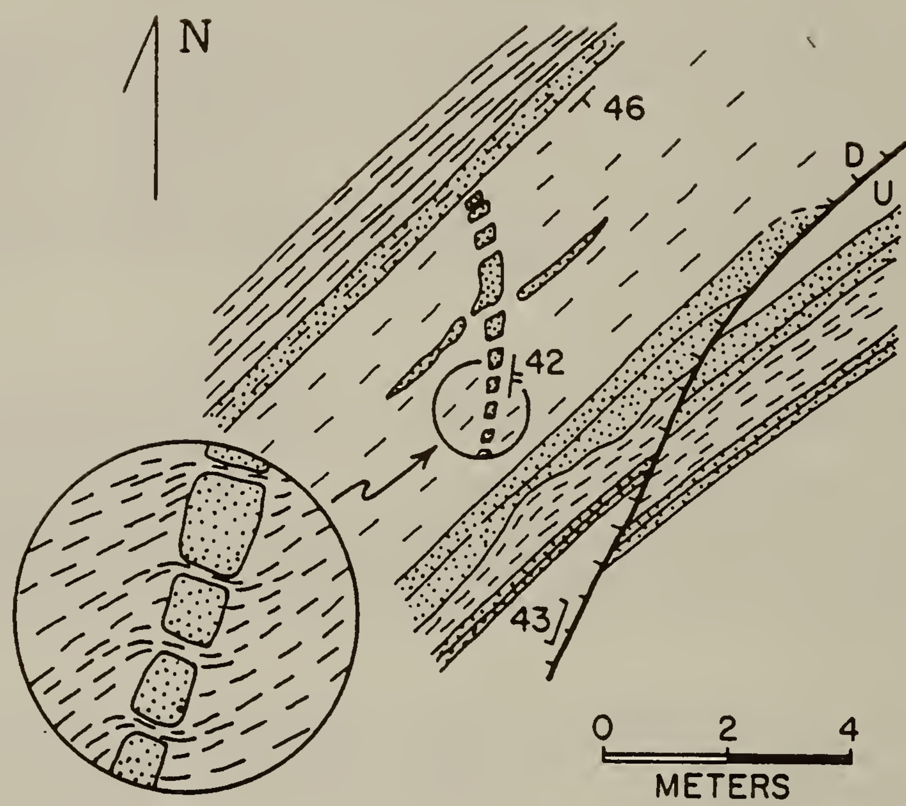



FIGURE 15. CLASTIC DIKE AT LOCATION #6

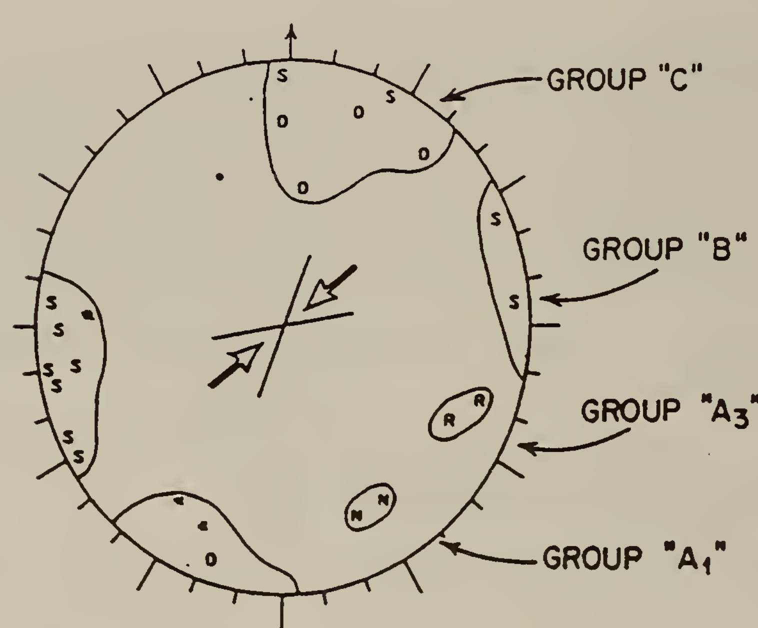



FIGURE 16. FAULT GROUPS AT LOCATIONS 7 AND 8

constricted between growing dike segments. The feature would appear to be the result of lateral injection within the mudstone bed. A likely scenario is fluidization by seismic shock of an unconsolidated sand which was prevented from dewatering by enclosure in the less permeable mudstone.

**LOCATION #7. Faulting in lake bed #0 and the sandstones of unit B**
-- Designation of the first lake bed here as #0 is in deference to a common terminology using #1, 2, and 3 for the overlying triplet of lake beds which comprise unit C. Bed #0, at the base of the resistant sandstones of unit B, is much less prominent and has much less organic matter than the overlying ones. In the island in midstream, it is even less prominent than on the shoreline exposures. However, it does contain carbonate concretions and some organic matter in a fine-grained matrix. Color changes along strike suggest oxidation or reduction reactions from circulating fluids

-- The lake bed has many small-scale bedding plane fault and fold structures. They are similar to but not so well exposed as those at location #10.

-- Several strike-slip faults cut the overlying sandstones and penetrate into the lake beds. Most are strike-slip but examples of curvature of slickenlines into more dip-slip displacement indicate thatlocal stress fields in the moving blocks permitted some transtensional motions. The plot (fig. 16) of the faults from this location up to the dam shows conjugate groups B and C with appropriate left and right-lateral displacement. (The opposite senses of motions indicated for a few of the slickenlines of the figure are probably the result of mis-interpretation of motion sense. The plot shows the original field determination of motion sense. The field data also include a certainity of motion value and these irregular displacements were designated as low certainity.) The faults are for the most part polished and little mineralized. Some interesting minor drag folding, crude cleavage development, and minor breccia production can be seen at the intersections or splays of some of the faults. Tracing of the faults upsection shows them disappearing into zones of more intense jointing.

-- About 25 meters downstream from the bridge is a relatively inconspicuous one meter-wide kink band which can be traced for about 20 meters with orientation N70E, 75 NW. It folds and hence postdates the N45E joint set. Formation of the kink requires compression within about 30 degrees of a prominent, pervasive anisotropy... bedding in this case. Similar kinks appear several hundred meters downstream of location #2 and locally have minor fault displacement associated.

**LOCATION #8. Lake bed triplet of unit C**
-- The prominent sandstone of unit B lies under the bridge. Lack of outcrop immediately above marks the largely covered trace of lake bed #1. At very low water levels it is partially exposed. The second and third lake beds are well exposed, the third being in the foot of the dam.

-- Abundant sedimentary features are present including dolomitic contcretions and infilled mudcracks (in the top of the third bed). Bedding plane faults and veins with fibrous calcite are present, similar to better exposed ones at location #10. Many minor folds and small faults are present.

-- Olsen (1986 and personal communication 1988) suggests that the tripletof lake beds represents climatic changes associated with Malenkovich cycles. Some lively discussion on the trip may arise concerning correlations of this triplet with others in the Mesozoic basins of the Appalachians.

**LOCATION #9. NE end of spillway island. Lake bed #3 and the spillway sandstone, unit D.**
-- At low water, access is possible by walking across sandstone ridges beneath bridge. For the trip, we will try to get permission to walk across the dam from Massachusetts Electric Company.

-- Geologic relationships are illustrated in the map of figure 4 and the cross sectional cartoon of figure 17. The dam foundations are in the resistant spillway sandstone. This unit also forms the resistant base for the north end of the dam and for the next bridge pier to the south.

-- Correlation with lake bed# 3 on the mainland is based on proper thickness, deep clastic-filled mudcracks, and septarian concretions comprised of dolomite (John Hubert and Paul Merieny, personal communication, 1988).

-- The "island" is a fault block splay of the Falls River Fault similar to the ones on the Canada Hill Fault (see photo, figure 10). The projection of lake bed #3 from previous location would pass considerably above bed #3 at

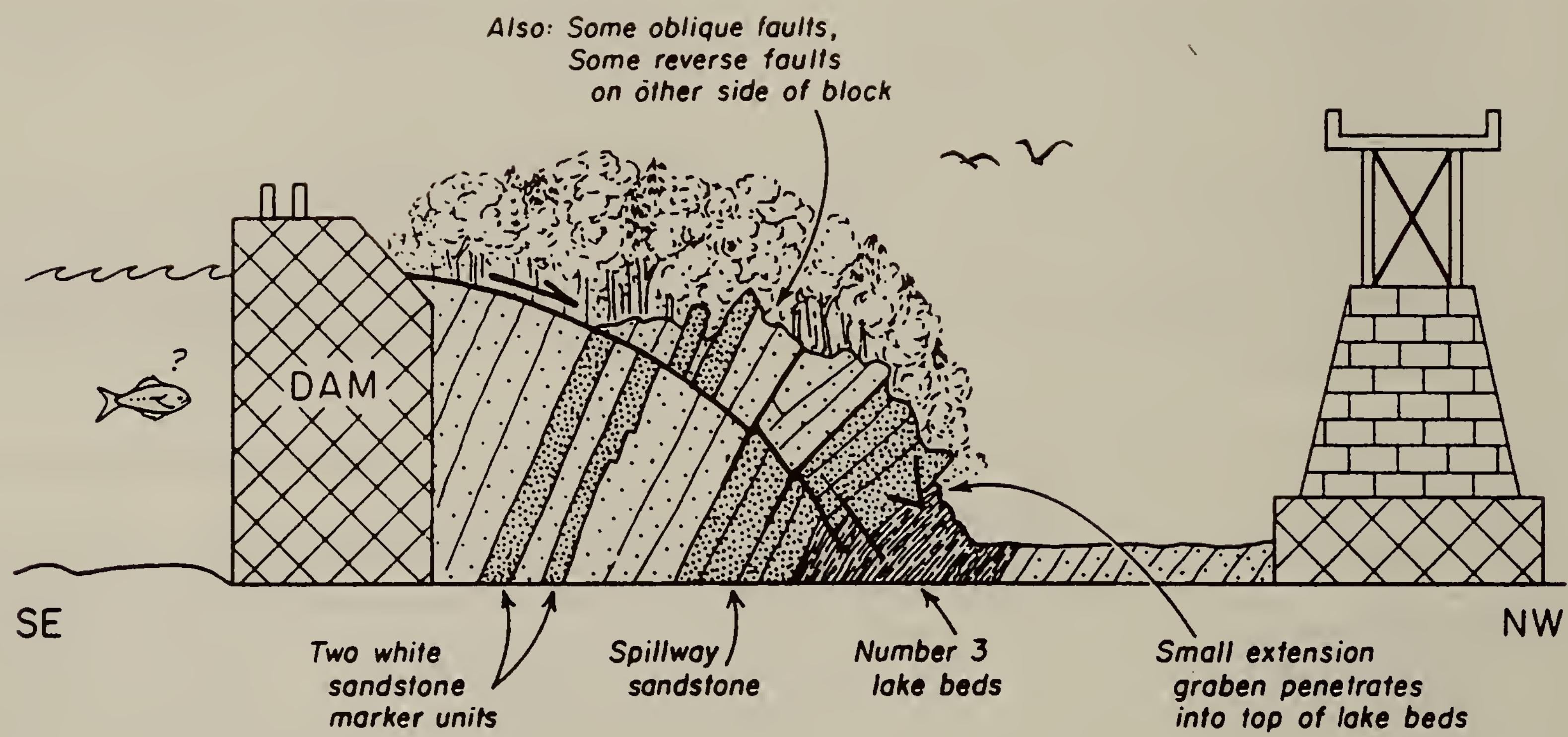



FIGURE 17. FAULT RELATIONS ON NE BLOCK OF SPILLWAY ISLAND (Location 9)

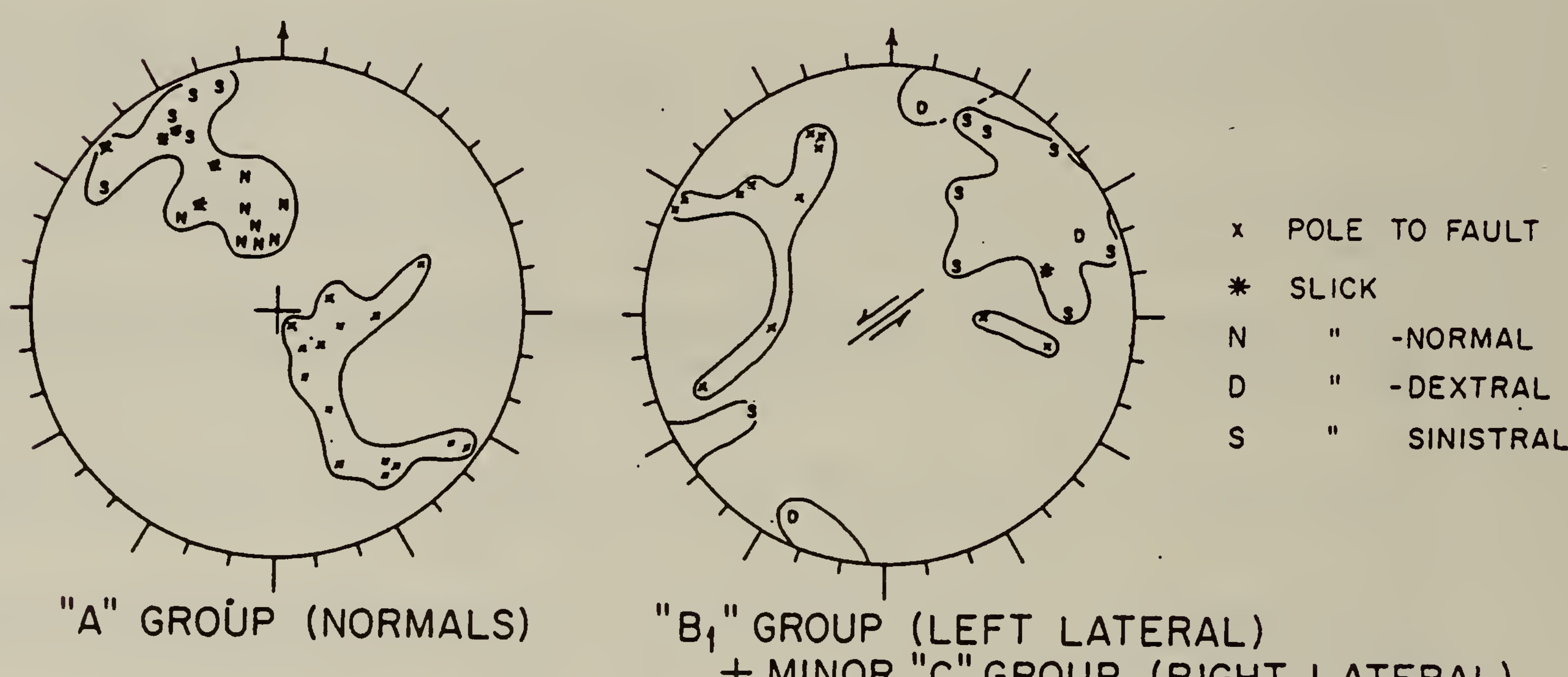



FIGURE 18. FAULTS AT LOCATION #9

this location. The fault must pass very close to the bridge pier and very close to the south end of the spillway gates. At extreme low water, the disturbed beds are visible next to the gates.

-- Much of the surface of the island is a large curving normal (?) fault surface (fig. 17). With present dips, the fault would be regarded as normal but it is essentially perpendicular to bedding, a paleo-vertical structure. At the contact with the underlying lake bed, the fault has others parallel to it, which bend and drag the lake beds. One small graben drops the overlying clastics about half a meter into the lake beds with the line of intersection of the conjugate faults at 215, 10 degrees.

-- Fault motions on the island are indicated on figure 18. Group A1, normal faults dominate but some probable revrse motions of group A3 type can be found, possibly a reflection of a younger compressional phase. Present but not so prominent are strike-slip and oblique faults of group C. A few group D faults are also present.

-- Toward the SW end of the "island" beds of the sandstone between lake beds #2 and #3 can be seen to curve and have increased amounts of displacement as they approach the open gut of water and the wall of rock beyond. The wall is composed of the resistant spillway sandstone dropped by the next and largest of the splays of the Falls River Fault (fig. 4).

**LOCATION #10. Lake bed #4. SW end of spillway "island."**
--Cross with care onto concrete apron of the dam. Largest of the Falls River Fault splays passes under concrete here (fig. 19). Immediately next to the main fault is a concentation of small A1,2, and 3 faults in the sandstone (right hand net of figure 20). Note breccia and mineralization.

--Lake bed #4 differs from any of those exposed on the mainland in having dolomitic laminites rare but present, a dolomitic concretionary sandstone at the base, an organic mudstone with flattened dolomitic concretions at the top and in having very high organic content with abundant hydrocarbons expelled into open fractures (John Hubert and Paul Merieny, personal communicatioin 1988). Fish scraps are abundant in a few locations.

--The sequence of structural events recorded in the pace and compass map of figure 19 and the net plots of figure 20 are given below.

(1) Early isoclinal folding with axes plunging shallowly at S30W. The best example of these folds is visible under the concrete at the SW end of the exposure. Cleavage is present in some of the axial plane regions of some of the folds but isoclinally folded calcite veins or other indications of mineralizing fluids are rare to absent. Fold asymmetry suggests top moving upward toward the NW. An early compressional phase seems likely in that the motion sense is opposite that expected for gravity drive in a SE tilting basin.

(2) Group F of bedding plane faults. These were associated with fibrous calcite growth with individual fibers plunging shallowly toward the NW. Slickenlines on the faults plunge shallowly and consistently at S70E. A system of thin fibrous veins is closely associated with and commonly merges into the faults. Displacement sense is difficult to determine from offset relationships but the extension direction of the calcite fibers clearly indicates a down to the SE or normal sense of displacement. Slow gravity sliding to the SE of the entire mass of basin fill is a reasonable model for the origin. Movements may have been complex as the fibrous veins, faults, and some thin beds are mildly folded and crinkled about axes plunging S20W.

The late stages of growth of the calcite fibers include some solid hydrocarbons indicating depth of burial and temperatures had reached the organic maturation stage.

(3) Extensive normal faulting common throughout much of the area is not common at this location except immediately next to the Falls River Splay. Development of well defined joints must have taken place in the well lithified sandstones and conglomerates because these structures are utilized for parasitic faulting in the next phase. The degree of lithification is indicated by the fracturing of joints through the pebbles of conglomerates.

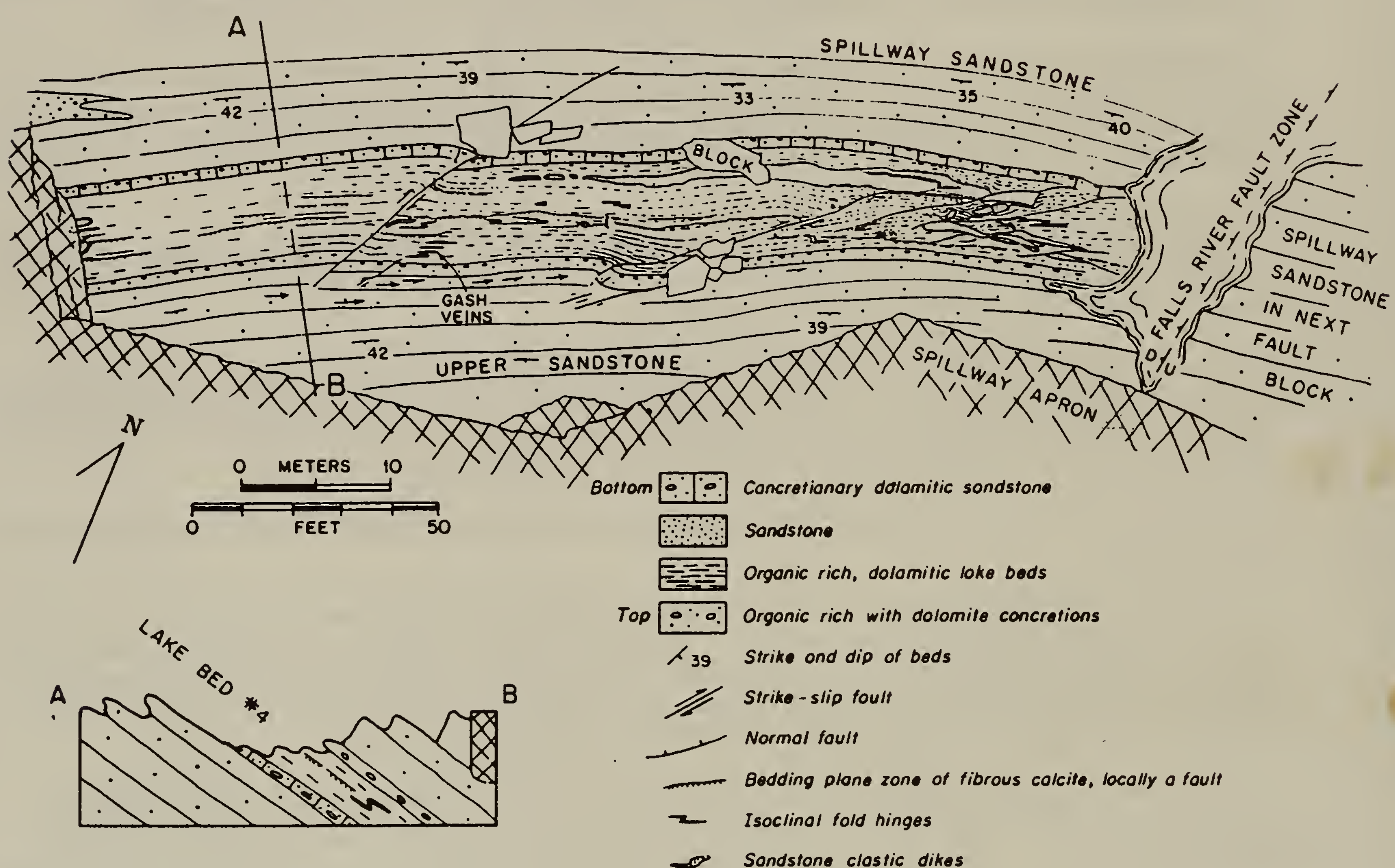
(4) Prominent conjugate strike-slip motion on both the B and C groups of faults. These motions are prominently displayed on the many large slickensided joint surfaces of the overlying sandstone and in faults crossing the lake beds (fig. 19). Associated dolomitic veins with en echelon arrangements show a properly oriented S50E sigma 3 to be in accord with a N40E sigma 1. The faults have calcite, dolomite, quartz, siderite, pyrite, chalcopyrite, galena mineralization. Hydrocarbon coatings on veins and open-space fillings are common, suggesting

this phase may have been at the height of organic matter maturation. Breccias are common and show evidence of multiple motions.

-- Jointing is well developed in the adjacent sands but rare to completely absent in the lake beds. This probably reflects the very slow lithification of the water-rich lake beds.

FIGURE 19.

LAKE BED #4, SOUTHWEST END OF SPILLWAY "ISLAND" (Location 10)



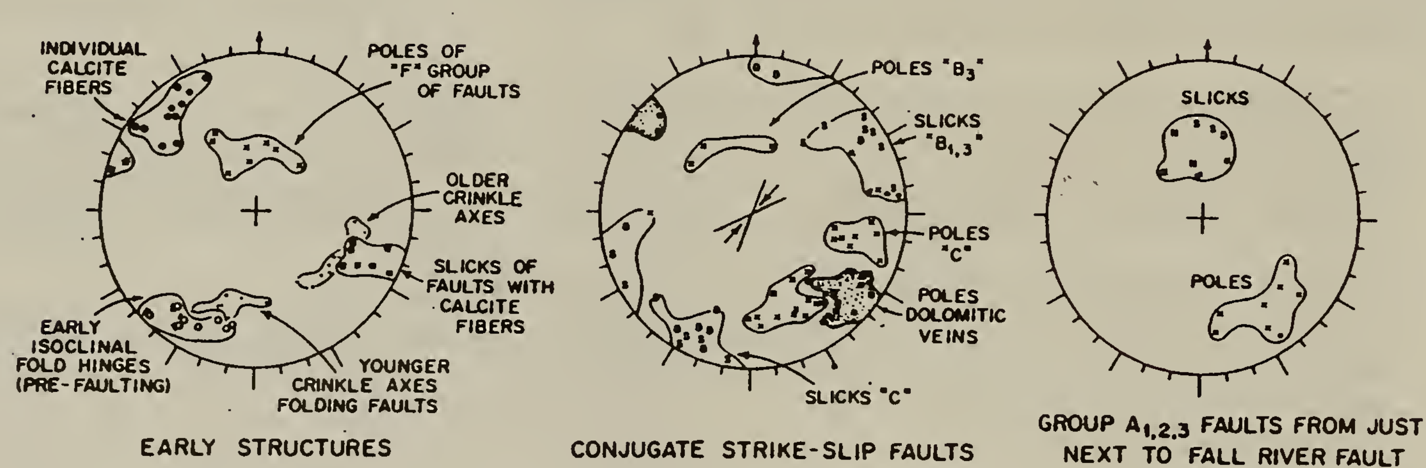




FIGURE 20. GEOMETRIC RELATIONSHIPS AT LOCATION #10